CONTINUOUS SEISMIC REFLECTION PROFILING IN THE STRAIT OF GEORGIA, BRITISH COLUMBIA

by

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ABSTRACT

Approximately 790 kilometers of continuous seismic reflection data were obtained with a 5000 joule Sparker in the Strait of Georgia, southwestern British Columbia. The Strait is a geological boundary between Upper Cretaceous Nanaimo Group rocks of the Vancouver Island area and Late Cretaceous-Early Tertiary continental rocks found in scattered outcrops on the southern mainland. Coast Intrusives form mountains on the mainland northeast of the Strait.

The Fraser River has built a large submarine delta across the Strait and is the main source of Recent sediments. Deposition is occurring mainly on the delta front and in deep basins to the northwest. In the basin adjacent to the delta, flat-lying bottomset beds average about 200 meters in thickness. An older layer of bottomset beds in this basin overlies bedrock and extends under the present foreset beds. Thinner sedimentary layers of possible hemipelagic origin overlie Pleistocene banks and ridges along the mainland north of the delta. No significant amounts of Recent sediment are presently accumulating in the Strait south of the delta. Erosion of possible Late Pleistocene deltaic sediments has deepened the Strait in that area.

Pleistocene deposits of probable drift, till and interglacial sediments occur mainly along the northeast side of the

Strait. One extensive stratified deposit, possibly correlated with exposed Pleistocene deposits on nearby shorelines, may reach 550 meters in thickness. Below the Pleistocene, stratified reflectors, suspected to be Late Cretaceous-Early Tertiary bedrock, unconformably overlie Coast Intrusive bedrock along the mainland shore. The reflectors dip seaward at 8 degrees or more.

Along the southwest Island coast Upper Cretaceous bedrock dips into the Strait. Deformation, most severe in the south, decreases northward. Dips of bedrock reflectors become less in mid-Strait before disappearing under delta deposits toward the mainland. Some synclinal and anticlinal folding occurs near mid-Strait.

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CHAPTER I

INTRODUCTION

I. PURPOSE AND EXTENT OF THE INVESTIGATION

With the advent of new tools of geophysics and geology it has become possible to study in detail the sub-sea structure off the continents and under the deep oceans. One such tool is the Continuous Seismic Profiler (abbreviated to CSP) (Ewing and Tirey, 1961; Hersey, 1963). This instrument operates on the same principle as a conventional ship-board echo sounder: intensive sound pulses are generated at a fixed repetition rate and echoes are recorded in the listening interval between pulses. It differs from echo sounders, however, in that operating frequencies are lower and sound energy is much greater in order to give significant penetration into the sediments and rock under the sea floor.

On the Pacific coast of Canada, no investigation into the use and value of the continuous seismic profiler was undertaken by a Canadian scientific institution prior to 1966, although exploration companies had used this instrument in British Columbia waters for at least three years previously. However, most of their results are confidential. Since the entire area of the British Columbia coast was

uninvestigated by this technique, at least in the name of research, it was desirable in a pilot project to determine the effectiveness of the instrument and to learn its characteristics and idiosyncracies in the channels and fiords as well as in more open waters of British Columbia. It was also desirable to combine this with a preliminary investigation of an area of geological importance. Thus the purpose of this investigation has been not only (1) to use and evaluate the continuous seismic profiler on this coast, but also (2) to use it in an attempt to solve problems of undersea geology. An additional objective has been to present the information gained on the use of CSP in such a way that geologists and non-technical personnel could profit from it.

The area of the investigation was, in part, determined by the time at which the equipment was available. It was possible to obtain a continuous seismic profiler only during the winter months. These months on the North Pacific coast are cold, foggy, and often stormy. Good seismic records are not made under these conditions. It was therefore decided to remain in the relatively sheltered waters of the Strait of Georgia between the mainland and Vancouver Island where, if gales did arise, the work could be continued in the inlets which run into the Strait.

Another important factor in the choice of location was the geological interest in the Strait of Georgia.

Several different geological provinces border on the Strait and their effects on the geomorphology of the Strait were not known. A variety of geological situations could occur under it. These could possibly be recorded and resolved by continuous seismic profiling. The Fraser River delta, for instance, extends into the Strait. Seismic profiles through this feature could help to understand its effect upon sedimentation in the area. Coast Range intrusives extend under the Strait from the northeast while Upper Cretaceous bedrock continues under the Strait from the southwest. Mesozoic and Paleozoic rocks also occur at various points around the perimeter. Pleistocene uplands are common in surrounding areas. It is desirable to know the distribution of these various bedrock and sediment units on the sea floor relative to each other. Studies in progress in adjacent inlets might also benefit by seismic profiles down those inlets.

Available ship time limited the area of study to that part of the central and southern Strait of Georgia extending from Ballenas Islands in the north to Patos Island in the south. This is an area of some 3770 square kilometers over which topography, relief and sedimentation vary considerably. Steep-sided and irregular ridges support near sea-level peaks while deep, flat-floored basins with water depths exceeding 420 meters occur nearby. Near the Fraser River Delta active sedimentation stretches into the Strait, while

in other areas relict surfaces of Pleistocene and pre-Pleistocene rock and sediment are exposed on the sea floor. Figure 1 shows the study area in relation to the whole of the Strait of Georgia and southwestern British Columbia.

II. EARLY HISTORY

History records the first European to visit the waters of the present Strait of Georgia as a Spaniard, Alferez Quimper who, in June of 1790, sailed through the Strait of Juan de Fuca and investigated its eastern end. Although he did not proceed far into or beyond the San Juan Islands, he reported an extensive tract of water lying to the northwest. One year later another Spaniard, Jose Maria Narvaez, under Lt. Francisco Eliza, entered the Strait of Georgia and explored as far north as 50°N latitude. Eliza named many of the islands and gave to the Strait the name 'Gran Canal de Nuestra Senora del Rosario la Marinera'. Fortunately this name did not survive. The following year, 1792, a chart of the southern Strait was drawn by the Spanish explorers Galiano and Valdes. At the same time, the English explorer and cartographer Captain George Vancouver entered the Strait in search of a 'northwest passage' through the North American continent. He also proceeded to chart the area, and in doing so, met the Spaniards near Point Grey on Burrard Peninsula. Vancouver's chart of the area shows the excellence of

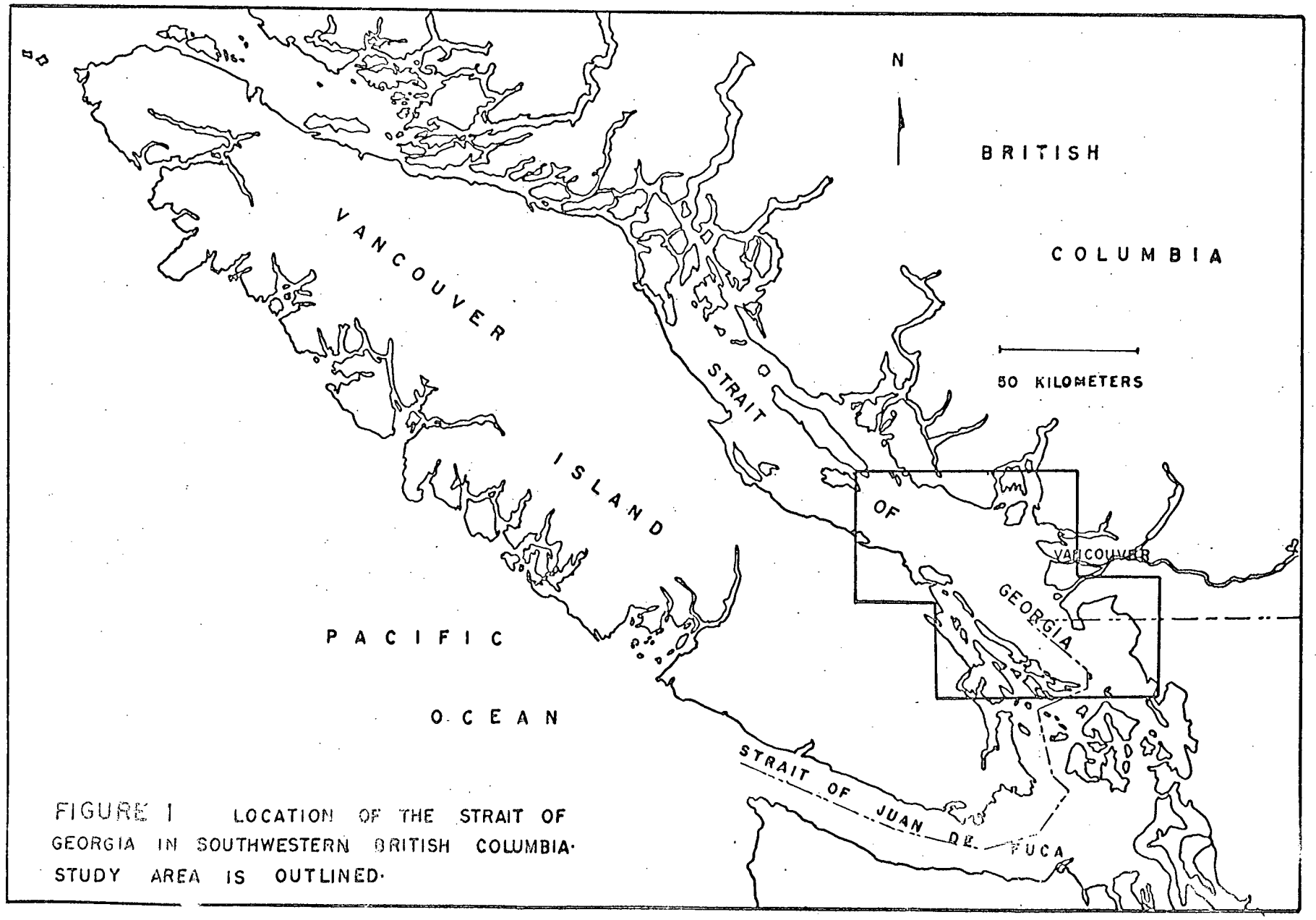


FIGURE 1 LOCATION OF THE STRAIT OF GEORGIA IN SOUTHWESTERN BRITISH COLUMBIA. STUDY AREA IS OUTLINED.

the work of this fine seaman. The complete area from the southern limits of Puget Sound to Smith's Inlet, north of Cape Caution, a linear distance of 670 kilometers, was included. With all inlets and fiords, a coastline of several thousand kilometers was surveyed in fine detail in just four months time.

In honour of his king, George III, Captain Vancouver gave the name 'Gulf of Georgia' to the extensive region from Queen Charlotte Strait to Puget Sound. From this comes the name 'Gulf Islands' given in the present day to the Canadian islands off the southeast coast of Vancouver Island. The 'Gulf of Georgia', since Vancouver's time, has been reduced in extent to the body of water bounded on the south by the San Juan Islands and on the north by Desolation Sound and Discovery Passage. The name on recent charts has also been changed to the Strait of Georgia. Neither Captain Vancouver nor the Spanish explorers discovered the Fraser River although the Spaniards in particular were looking for such a feature, having heard from the Indians that a large river existed. Vancouver twice crossed the delta front no more than eight kilometers from the river mouth. Obviously not knowing the physiography of deltas, and in spite of abundant evidence in the form of logs and stumps aground on the mud flats, he did not recognize that a river discharged there. In fact, he emphatically denied the possibility of the river to the

Spaniards upon their meeting at Point Grey in 1792, probably not more than six kilometers from the north arm of that river! It remained for Simon Fraser in 1808 to discover the mouth of the river by following it from inland. Fraser, however, was beaten back from the river mouth by hostile Indians of the Musqueam tribe before he could explore further in the region.

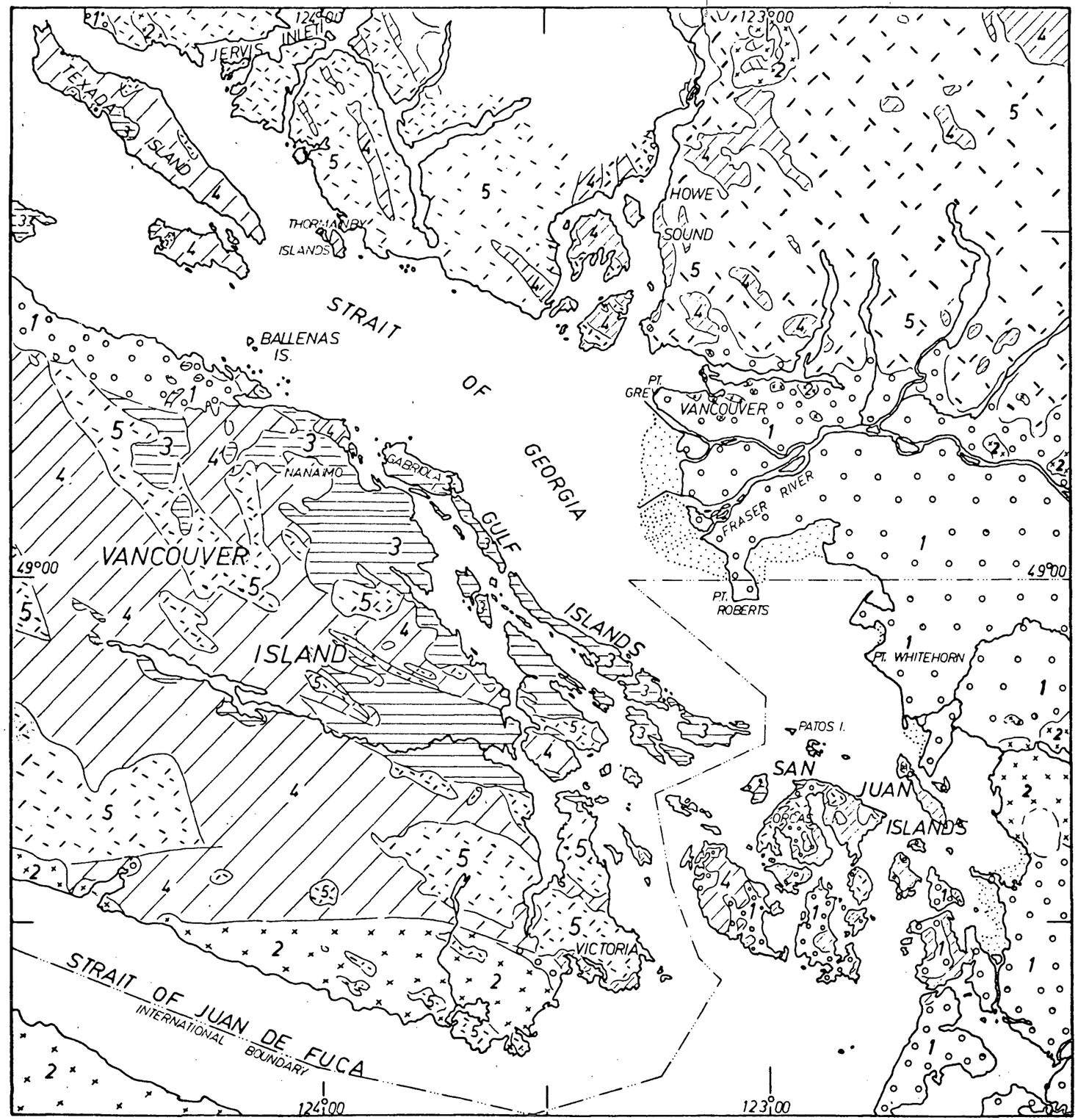
III. REGIONAL GEOLOGY

The Strait of Georgia, or colloquially, Georgia Strait, is a semi-enclosed body of tidal marine water separating southern Vancouver Island from the mainland of British Columbia (Figure 1). The Strait extends 232 kilometers in a northwest-southeast direction, from the American San Juan Islands in the south to Quadra Island in the north. The width varies from seventeen and one half kilometers between Texada Island and Vancouver Island, to thirty-five kilometers between Point Grey, near Vancouver, and Valdes Island in the Gulf Islands. The average width is approximately twenty-eight kilometers.

The Strait of Georgia is a submerged portion of the Coastal Trough, a regional topographical low extending from Puget Sound in Washington, USA, to Dixon Entrance, north of the Queen Charlotte Islands (Holland, 1964, p. 32). This topographic low, known in the Strait of Georgia region as the

Georgia Depression, is surrounded on each side by high mountains and underlain by thousands of meters of late Cretaceous, Tertiary and Quaternary sediments. The Strait itself marks a boundary between exposures of important geological formations of the Georgia Depression. Figure 2 describes the regional geology of the area surrounding the southern Strait of Georgia.

Along the southeastern coast of Vancouver Island, including the associated offshore Gulf Islands, a strip of low-lying country forms an unsubmerged part of the Georgia Depression. It is largely underlain by Upper Cretaceous marine sedimentary rocks of the Nanaimo Group (Clapp, 1912, 1913, 1914; Usher, 1952). North of the town of Nanaimo an arch of older volcanic and intrusive rocks is exposed on Vancouver Island which appears to separate the lowland into two basins, the Nanaimo Basin in the south and the Comox Basin in the north (Figure 2). Across the Strait on the mainland, no outcrops of Nanaimo Group sedimentary rocks occur. Instead, the Coast Range north of Burrard Inlet forms high mountains which rise above a narrow lowland strip along the Strait. Where it exists, this lowland, called the Georgia Lowland, is underlain mainly by granitic rocks of the Coast Range intrusives (LeRoy, 1908; Bacon, 1957; Mathews, 1958; Phemister, 1945; Armstrong, 1960 a; Roddick, 1965). Older rocks occur sporadically along the mainland



LEGEND

1 — RECENT and PLEISTOCENE
Alluvium, drift, interglacial sediments, etc.

2 — TERTIARY SEDIMENTS and VOLCANICS
Includes Chuckanut, Kitsilano, Burrard and Sooke Formations, Metchosin Volcanics and Garibaldi Group.

3 — UPPER CRETACEOUS SEDIMENTS
Mainly Nanaimo Group.

4 — PRE-UPPER CRETACEOUS SEDIMENTS, METASEDIMENTS, IGNEOUS and VOLCANIC ROCKS. Includes Vancouver Group, Bowen Island Group, Jarvis Group, Leech River Formation, Sicker Group, Gambier Group, San Juan Group, Malahat Volcanics.

5 — COAST RANGE INTRUSIONS
and post-tectonic plutons of Vancouver Island and elsewhere.

Compiled from:

Geological Survey of Canada, Map 1069A, Victoria-Vancouver Sheet, 1957; Map 49-1963, Alberni Area; Geologic Map of British Columbia, 1963.

J.E. Muller, G.S.C., unpublished maps, 1967.

Geologic Map of a part of Southwestern British Columbia and adjacent Washington. From Geological Society Field Trips, 1960. Published by Vancouver Discussion Club.

Washington Geological Survey, Map of San Juan Islands

FIGURE 2. Distribution of Bedrock Geology in the Region of the Southern Strait of Georgia, British Columbia and Washington.

north and northwest of Burrard Inlet. South of Burrard Inlet, a re-entrant, occupied mainly by the lower Fraser River valley, is almost encompassed by mountains. Called the Fraser Lowlands, the re-entrant is open only to the west where the Fraser River delta is building into the Strait of Georgia. Tertiary rocks dip off the mountain front into the 'Whatcom Basin' (Newcomb, et al, 1949; Hopkins, 1966) which underlies the Fraser Lowland between Vancouver and Bellingham. Outcrops of these rocks occur at several points on the north and south shore of Burrard Inlet and at various locations surrounding the Fraser Lowlands (Figure 2). Under the Tertiary fill of the Whatcom Basin, deep wells (Richfield Pure Sunnyside; Richfield Pure Pt. Roberts) reveal the presence of thick sections of Cretaceous sedimentary rocks of continental origin indicating that the Whatcom Basin may be, in part, analagous to the Nanaimo Basin. These sediments probably overlay deeper granitic basement (White and Savage, 1965).

No Tertiary rocks similar to those found in the Whatcom Basin occur on the eastern half of Vancouver Island. In fact, the known Tertiary sedimentary rocks on Vancouver Island are outcrops of limited extent along Juan de Fuca Strait and the western coast of the island. However, the upper part of the Gabriola Formation of the Nanaimo Group is not fossiliferous and could be Early Tertiary in age (J.E. Muller, Geological Survey of Canada, personal communication).

The Strait of Georgia region has undergone several episodes of glaciation during which ice filled the Strait to a depth of 1,500 meters (Glacial Map of Canada, 1958). Subsidence of the land relative to the sea due to ice loading was as much as 230 meters or more (Armstrong and Brown, 1954). Ice sculptured rock surfaces and left in its path varied deposits of drift and till. At least two glacial episodes have been noted on Vancouver Island (Fyles, 1963) and at least three major glacial advances are known from the Vancouver area (Armstrong, 1956, p. 4). A late readvance of Cordilleran ice into the Fraser Lowland, the Sumas glaciation, resulted in a valley glacier which, although not extending to the present day Strait of Georgia, deposited drift in the Fraser Lowland (Armstrong, 1960 b). Between glacial episodes, non-glacial periods brought erosion and deposition, the effects of which are shown over wide areas about the Strait. Thick interglacial sediments are common on several islands in the north-eastern part of the Strait (Bancroft, 1913; McConnell, 1914) as well as in the Fraser Lowland (Armstrong and Brown, 1954; Johnston, 1923; Easterbrook, 1963) and on Vancouver Island's east coast, mainly northwest of Nanoose Bay (Fyles, 1963; Halstead and Treichel, 1966). The terminology of Armstrong, et al, (1965) will be used throughout this thesis in naming Late Pleistocene events.

The Fraser River, with a mean annual discharge of 28,500 cubic meters per second, is by far the most important river flowing into Georgia Strait. Other rivers, of lesser importance, mainly discharge into the heads of long inlets which connect to the Strait and, except for the Fraser, only minor streams enter directly into the Strait.

Since Pleistocene time, the Fraser River has built a large sub-aerial and submarine delta into the Strait of Georgia. The sub-aerial part extends from near the city of New Westminster to the sea, a distance of twenty-four kilometers. Tidal flats up to eight kilometers wide occur off the sub-aerial part. Beyond these, the submarine delta front continues for another fifteen to twenty kilometers under the Strait. It is estimated that the present delta has required less than 11,000 years but more than 7,300 years to build to its present state (Mathews and Shepard, 1962). It is now advancing at an average rate of eight and one half meters per year into the Strait at the ninety meter level (op. cit.).

IV. PREVIOUS WORK IN THE STRAIT OF GEORGIA

Previous Geological Studies

Interest in the geology of the area about the Strait of Georgia began shortly after the establishment of the first community at the present site of Victoria in 1843. Indians, observing the settlers interest in 'black stones', or coal,

pointed out the presence of a small deposit near Beaver Harbour on northern Vancouver Island. This was worked for a short time in 1849 before being replaced by development of the Nanaimo coal workings in 1852, also discovered initially by Indians. Thus the first geology of the area was done by Indians.

In 1856, the discovery of gold along the Fraser River and its tributaries attracted a great influx of prospectors and miners to the area. When the known gold deposits were depleted many prospectors searched the coastal areas for precious minerals as indicated by mineral claims in the area dating back to the very early years.

The prospect of mineral resources brought the land areas to the early attention of professional geologists. However, it was not until 1921 that a marine geological investigation in the Strait of Georgia was published. In that year, Johnston (1921) reported on sediments of the Fraser Delta including the river mouth and adjacent Strait.

Waldichuk (1954) gave a generalized account of the character of bottom sediments in the Strait of Georgia using one hundred bottom samples and data from the hydrographic charts. His work showed surficial sediments consist mainly of soft, silty clays and clays, with local sand and gravel patches implying the presence of non-depositional sites.

This information was later incorporated into a larger volume (Waldichuk, 1957) on the general oceanography of the region.

Mathews and Shepard (1962) conducted a hydrographic and bottom sampling program off the Fraser River delta between Point Grey and Point Roberts. They showed sandy sediments predominate to the south of the present river mouth whereas silt predominates to the north. The absence of muds and the origin of the sands were attributed to checking of coarse bottom transported sediments by the flood tide entering the river. With the south-setting ebb tide, the coarser materials would be released and distributed to the south. Another explanation, of relict sands, although an admitted possibility, was thought to be less likely. A topographic ridge to the northwest of the main river mouth, called Fraser Ridge in this thesis, was suspected to have a Tertiary or Cretaceous core. Sampling indicated soft mud covered the ridge (op. cit., p. 1425) although bottom currents are high south of the river (Pickard, 1956). Their data also showed anamalous hills with a relief of twenty or more meters occurred on the delta front off the river mouth. Mayers (1968), utilizing some of the CSP data of this study, did further studies on this hill topography.

Cockbain (1963a) studied the area from Sand Heads to Ballenas Islands using a Precision Depth Recorder (PDR) and a 12 kiloHertz Edo echo sounder. The 900 kilometers of echo

sounding track obtained were used to divide the area into a number of physiographic sub-divisions, some of which are retained in this thesis. Penetration of the sound waves of up to thirty-five meters into the sediments was obtained in favourable areas. These records exhibited layering in the unconsolidated sediments under the flat floors of the basins. Turbidity currents were suggested as a possible mechanism of sediment transport to these sites. Cockbain concluded that topographical differences between the mainland and Vancouver Island possibly reflect the different geology of the areas. He placed the boundary between structural regions at the boundary between submerged topographical areas. Cockbain (1963b) also described briefly the general sediment distribution in the Strait as a by-product of his research on the foraminifera of the region.

Previous Geophysical Studies

Little geophysical work has been done in the Strait. Milne and White (1960) published the first deep crustal refraction seismic survey of the region including the Strait near Vancouver Island. Later, White (1962) conducted further refraction studies along a profile down the northwestern side of the Strait of Georgia. Average P-wave velocities were computed for a two and a three layer model. He also reported on gravity measurements along the east coast of Vancouver Island.

A single gravity track has been recorded along the axis of Georgia Strait (Delingher, et al, 1966). Walcott (1967) recorded many additional gravity stations in the area. All previous gravity data have been compiled into a Bouguer anomaly map of western British Columbia (Stacey, et al, 1969).

White and Savage (1965) published results of refraction work including unpublished work mentioned in the above references along with additional data.

Tiffin and Murray (1966) gave a preliminary report on the results of the present study indicating the structural significance of the work.

Tseng (1968) used new analytical techniques (time-term methods) to amplify, improve and enlarge upon the seismic data obtained by Milne and White (1960) and White (1962). It is shown that in the southern Strait of Georgia there is a marked thickening of low velocity sediments with average P-wave velocity of 4.5 kilometers per second.

CHAPTER II

CONTINUOUS SEISMIC PROFILING TECHNIQUE

I. ADVANTAGES OF CSP METHODS COMPARED WITH HIGH EXPLOSIVE METHODS

Continuous seismic reflection profiles, that is, seismic data records which, when laid side by side, make up a composite and continuous record, can be obtained by conventional explosive seismic techniques, but it is difficult and dangerous. Large quantities of explosives must be loaded and carried aboard ship. At sea, consecutive explosive charges are required to be detonated in the water, generally at close intervals of time. This has been done (Ewing and Tirey, 1961; Officer, 1955; Paitson, et al, 1964; Savit, et al, 1964) but it is difficult to perform and even more difficult to maintain for long periods of time without undue strain and hazard to personnel, not to mention the high cost of explosives. These factors severely limited the use of continuous profiling techniques before the advent of a non-explosive seismic source.

Non-explosive continuous seismic profiling equipment was developed in the late 1950's (McLure, et al, 1958; Beckman, et al, 1959; Hersey, et al, 1961; Shor, et al, 1963). At that time graphic recorders were in common use with echo sounders and it was apparent on these records that echo sound-

ing could penetrate the sea floor to provide profiles of shallow underlying bedrock (Smith, et al, 1954; Smith, 1958). However, echo sounding equipment was neither sufficiently powerful nor able to operate at low enough frequencies to be useful in seismic operations. Consequently, the CSP method had to rely upon different techniques to create a sound pulse, principally either electrical discharge, gas explosion or discharge of air under water.

It is possible to generate a constantly repeatable discharge in the water every few seconds with any of these techniques. The number of operators required is small compared to a conventional seismic crew and very little hazard exists to personnel. One person can attend the equipment for hours at a time without fatigue. An additional great advantage is that only one ship is required and this does not have to be a large one. The methods of producing acoustical pulses without high explosives also enjoy the advantage of creating no ill effects to fish. They may therefore be used where explosives are prohibited by fisheries protection laws.

In high explosive seismic work, noise and reverberation generated in the sea water by the explosion and bubble pulse may obliterate arrivals from sub-bottom structure. With the non-explosive profiling equipment, a less energetic sound wave generates much less noise and reverberation and primary echoes from sea bottom and immediate sub-bottom reflectors

are usually distinct on the record. Because the sound pulse may be generated many times per minute, reflectors of limited areal extent or structural features of a small scale can reflect several sound pulses during the time of a traverse. The chance of detecting and correctly identifying such features is therefore increased many-fold. A survey ship commonly travels at a speed of six knots or more and fires at a repetition rate of four seconds or less. A sound pulse is thus emitted about every forty feet of travel. With such a shot density there is comparatively little chance of any significant shallow geological structure escaping detection. Although each outgoing pulse contains less energy than that produced by high explosives, the greater number of reflections, even though weak, permit a more reliable interpretation of otherwise questionable features. Even where the signal-to-noise ratio is low, coherence on the record of a large number of in-phase reflections allows them to be detected against the background of random noise. Thus the greater density of information as well as its greater detail on a CSP record often makes interpretation simpler and more certain than on most explosive reflection seismic records.

The main drawback to the CSP method has been its low energy output which, with higher frequency content, limits its penetration capabilities. This is presently being remedied in more expensive equipment. It is now possible to

shoot CSP surveys using very high power. For instance, 160,000 joule 'Sparker' reflection seismic equipment has been used for ocean surveys by private exploration companies (World Petroleum, March, 1968). A 2,000 cubic inch air gun has been used successfully in crustal refraction studies by scientists of Atlantic Oceanographic Laboratory, Dartmouth, Nova Scotia, over shot-detector distances exceeding one hundred kilometers. This gun can also be used for continuous reflection profiling. The energies and spectrums of these sources are equivalent to that of several pounds of explosives.

It should not be assumed, however, that CSP is displacing explosive seismic surveys. On the contrary, they are complimentary. CSP methods are useful to delineate near surface features in preliminary surveys. Large areas can be covered in relatively short time. Smaller areas can then be intelligently selected for more costly detailed high explosive deep seismic reflection studies. Even here, CSP is useful to define the short-range sub-surface often lost in the deeper penetrating lower frequency high explosive work. The higher frequency output of the CSP adds to its ability to detect this detail. Of course CSP, as described here, can only be used in areas covered by water.

Continuous seismic profiling equipment does not, by itself, give information on seismic velocities. Wide angle reflection or refraction profiles from which this information

is calculated can be obtained (Le Pichon, et al, 1968; Houtz, et al, 1968) with additional equipment such as sono-buoys and radio receivers. Thus seismic refraction shots are still a useful part of a survey operation.

II. DESCRIPTION OF THE CONTINUOUS SEISMIC PROFILING EQUIPMENT

The continuous seismic profiling equipment used in this study was a 5,000 joule 'Sparker' manufactured by Edgerton, Germeshausen and Grier, on loan to UBC for the month of January, 1966, by Dr. M. J. Keen, Department of Geology, Dalhousie University, Halifax, Nova Scotia. It is called 'Sparker' because it makes use of an electrical discharge or spark to create a sound pulse in the water (Hersey, et al, 1961).

The sparker system can be described in two sections: signal generating equipment, and receiving and recording equipment. A block diagram of the major components of both sections is shown in Figure 3.

The signal generating equipment consists of power supplies, capacitor banks, a triggering device and a transducer. Electrical energy at 110 or 220 volts A-C from the ship's mains or from an auxiliary generator is transformed to four thousand volts and rectified in the power supply units. The high voltage D-C energy is then stored in

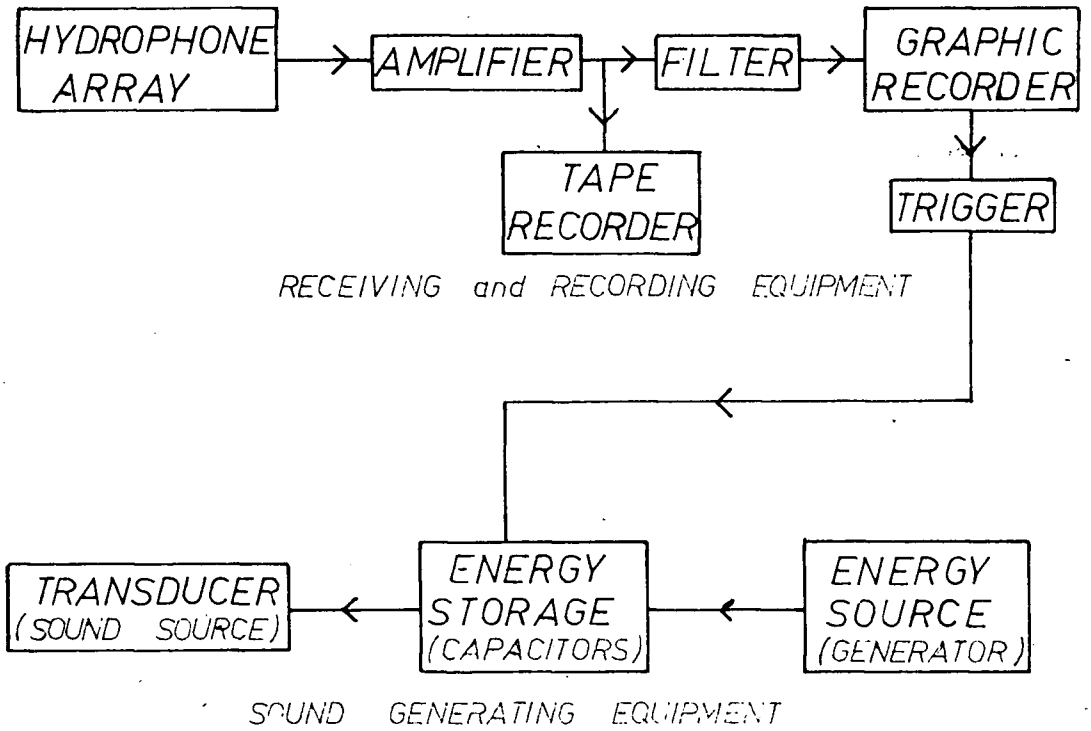


FIGURE 3. Block diagram of continuous seismic profiler.

capacitors until the instant of firing. An air gap switch between the capacitor banks and the transducer prevents premature leakage. The latter is trailed in the water over the stern of the ship. On command from the recorder, a trigger signal initiates the breakdown of the air gap by ionizing a path across it. Once in existence, a heavy current arcs across the gap until the capacitors are discharged to the point where they can no longer support the arc. Since this occurs within milliseconds, peak current flow to the transducer reaches several hundred amperes. The heavy conductor required for the passage of these high currents also serves as a towing cable for the transducer.

The transducer is essentially an open-ended cable which permits the current to be short-circuited through the seawater to a nearby ground return. In order to provide a low impedance short-circuit path, the cable divides at the transducer into three ends, each of which are brought to a replaceable spark-tip in a seven foot linear array called a sparkarray. The large energy pulse through the seawater causes the water around the spark-tips to vapourize with explosive rapidity, creating an intensive sound wave with essentially omni-directional characteristics. The signal generating equipment is shown in operation in Figure 4.

Since energy $E = \frac{1}{2} CV^2$ where E is in joules, C is capacitance in farads and V is voltage, it is apparent that

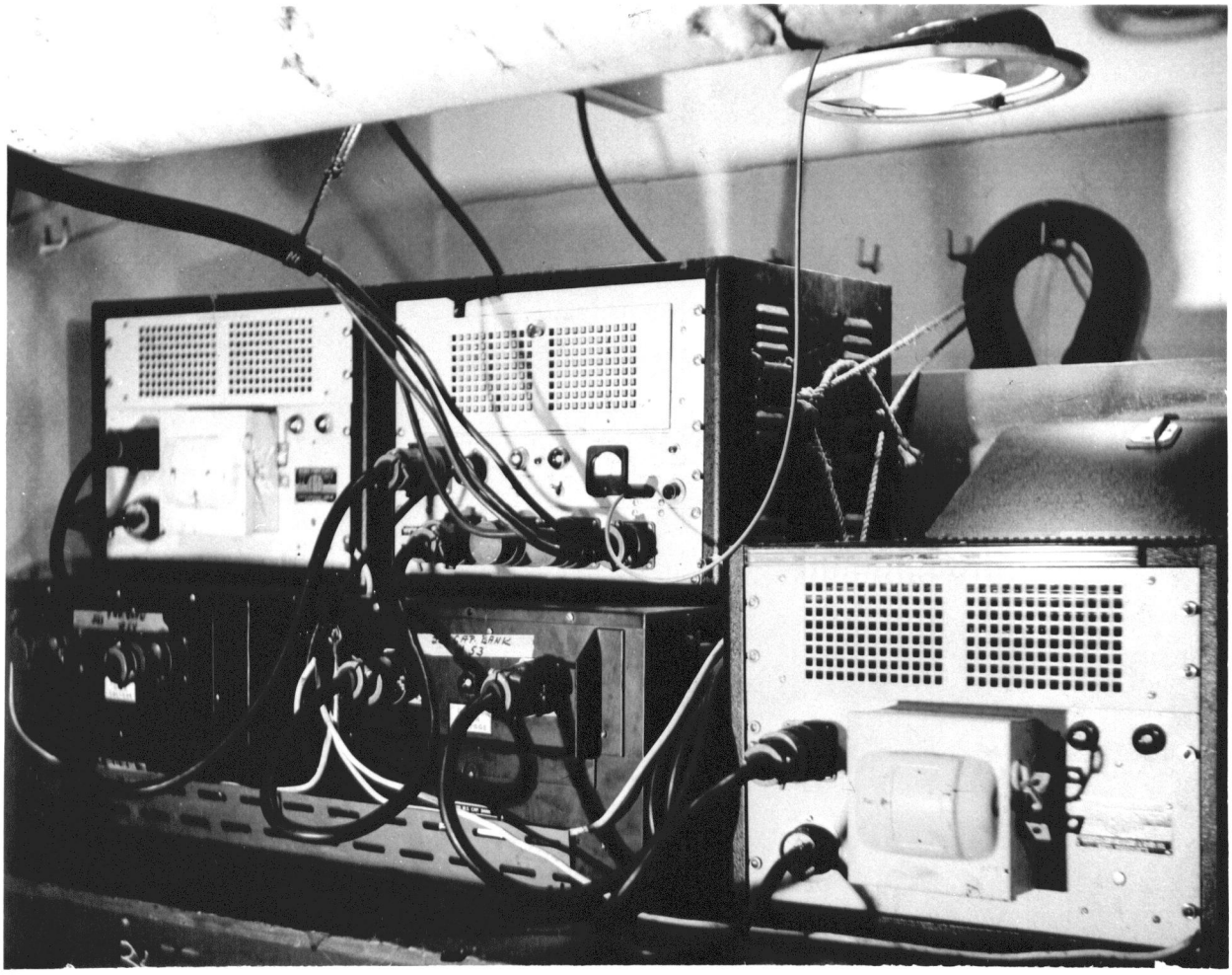


FIGURE 4. Signal generating equipment installed aboard CNAV Saint Anthony. Equipment consists of two identical power supply units at upper left and lower right, two capacitor banks at lower left and a trigger unit, top centre. A heavy four conductor cable leads from the trigger unit to the sparkarray.

by increasing either voltage or capacitance, or both, the amount of stored energy can be varied up to the limits set by breakdown of the physical components of the system. The equipment used for the present survey maintained voltage at a constant 4,000 volts and discrete energy variations were obtained by adding or removing banks of capacitors.

Echoes from bottom and sub-bottom reflectors are received by arrays of hydrophones towed in the water, also astern of the ship. For most of this survey, a short, linear untapered array of 10 hydrophones, connected in a series-parallel arrangement, was used. Later, additional survey lines were recorded using a 20-element 100-foot array giving a better signal-to-noise ratio. Signals from the array, amplified by a GeoSpace Model SA-216 seismic amplifier system, were filtered to reduce undesirable noise and recorded on chemically treated paper on an Alden 419 Precision Graphic Recorder.

III. SHIPBOARD INSTALLATION

Seismic data was obtained in two cruises. The first cruise, during which most of the data was collected, took place in January, 1966 aboard the tug CNAV 'Saint Anthony', a Canadian Naval Auxiliary deep-sea salvage tug.

The continuous seismic profiling equipment was received aboard ship just prior to New Years Day, 1966 and

was stowed comfortably in a small dry cargo hold below the after deck. Transducer and hydrophone cables were led up through a hatch cover and over the stern, one on each side of the ship. The low after deck was often awash, but the equipment was safe and working conditions comfortable, although getting to and from the temporary lab sometimes presented problems.

During this cruise, profiles were run in the Strait of Georgia, Howe Sound, Jervis Inlet and Bute Inlet, the latter two inlets being to the north of the study area. Winds during the survey ranged from calm to gale force but no time was lost due to weather.

A second cruise in June, 1967, made use of a small thirty foot aluminum-hulled launch manned by two people. Sparker equipment for this cruise was supplied by Dr. Gene Rusnak, United States Geological Survey. The equipment, except for recorder and amplifier, was housed in an open after cockpit and covered with a tarpaulin. A portable 5 kilowatt electric generator supplied operating power. The recording and amplifying equipment was stored in the small cabin and run from the ship's generator. Electrical noise through the aluminum hull presented a problem which was finally overcome by reducing the seismic energy output to 1,000 joules. Even at this low output, penetration obtained was almost as good

as with the full 5,000 joule output (compare Plate VII with Plate VIII).

Additional profiles were obtained with the use of this launch in the Strait of Georgia, Howe Sound and, unsuccessfully, in Pitt Lake, a tidal lake off the Fraser River. The latter was not successful because the natural salinity was insufficient to support electrical conduction across the transducer and an artificial saline environment could not be maintained around it while being towed through the water.

IV. NAVIGATION

During the survey positions were obtained by three point radar and polaris compass bearings by the ship's navigating officers. Reliability, which varies with factors such as distance from target, type of target and crossing angle of the bearings, was weighted at the discretion of the operator by a system of three numbers, number one being most reliable. In most cases, positions are assumed to be correct to within $\frac{1}{2}$ kilometer ($\frac{1}{4}$ nautical mile) or less. Positions in some areas where good radar targets were lacking, such as off the Fraser Delta, may be less reliable.

V. THE CONTINUOUS SEISMIC RECORD

The record generated by a precision graphic recorder as used in continuous seismic profiling is distinctly differ-

ent from the type of record usually obtained by other seismic methods. Signal processing is accomplished during the recording so that the final record is presented almost at the instant it is received. The procedure is repetitous, each shot being followed by a listening interval of one or two seconds then another shot. Since the record paper is moving out of the recorder, each individual shot is printed beside the previous one. The record is therefore continuously produced as the survey proceeds, and provides up-to-date information on the sub-bottom structure being traversed.

In order to appreciate the amount and type of information on the record, it is necessary to understand the basic principles involved in the reflection of sound waves through earth and water, and their collection, processing and presentation on a graphic recorder.

Paper Type

The specific graphic recorder used with the CSP equipment determines whether the record will be produced on wet or dry chemically treated paper. These are not interchangeable on any one recorder. The Alden Precision Graphic Recorder or PGR (Luskin, et al, 1954; Knott, 1962) used in this study requires wet paper. Wet paper has more shrinkage and a greater tendency to distort than the dry paper type. This can be troublesome as the paper refuses to lie flat.

Timing marks printed at the time of recording distort with the paper and preserve the integrity of the time domain. Travel times can therefore be accurately read from any part of the record. Wet paper has a wide dynamic range of tone contrast permitting good resolution between strong and weak signals. Marking is accomplished by electrical impulses through a wire helix wound on a rotating cylinder under the record paper. The helix has a pitch equal to the width of the record. As the helix revolves, at a speed determined by the vertical scale desired, the contact point scans at a constant speed across the paper. Echoes, timing marks or extraneous noise impulses travel from the helix through the chemically-treated paper to a second electrode on top, leaving a mark on the paper. The firing trigger to initiate the outgoing sound pulse is actuated by a mechanical switch at the moment the left hand end of the helix returns to the paper. Thus the output pulse is synchronized to the zero time mark on the record.

Timing Marks

The recorder is designed to produce a series of crystal controlled timing marks at set intervals across the record. The first heavy mark is manually adjusted to occur a few milliseconds prior to the trigger signal to compensate for towing depths of transducer and hydrophones. This mark

then corresponds to the water surface; that is, zero meters or zero time. In the accompanying plates, lighter timing marks can be seen at every 50 milliseconds (or 20-fathom interval at a velocity of sound in sea water of 4,800 feet per second). At each 250 milliseconds (or 100-fathom interval), the timing mark is heavier to facilitate reading the record. As the paper advances out of the recorder, continuous timing lines are produced on the record.

To provide a time scale in the horizontal direction as well as the vertical, the timing marks are automatically occulted for a period of approximately thirty seconds every five minutes. These small breaks are not easily seen in the record reproductions.

Signal Processing

Signal processing can take many forms depending on the type of information required. Various techniques for multiple reduction and signal enhancement exist using sophisticated equipment but the simplest and most direct signal processing is often adequate in the field. If the raw information is tape recorded, sophisticated methods may await a more favourable arrangement in the laboratory or computer.

Since noise is a problem on ships where electrical generators and motors as well as propulsion machinery are

operating, it is usual to filter the incoming signals in an effort to increase the signal-to-noise ratio. For instance, if the outgoing pulse does not contain frequencies lower than 80 Hertz (Hz) one can filter those frequencies and remove low frequency noise without loss of signal. Much ship noise is generated at 60 Hz which can be effectively removed by this method or by a notch filter at 60 Hz. Similarly, useful high frequency information is limited. For frequencies above a few hundred Hertz penetration is not great in the sediments. By filtering these high frequencies, noise can be considerably reduced. In practice one must usually trade off some signal for a gain in the signal-to-noise ratio. A one to two octave band-width is usually found to be sufficient to obtain good information with reasonably low noise.

As well as active or passive filter elements, other components are effectively acting as filters as well. The hydrophones may respond only over a limited range of audio frequencies. The amplifiers may also have a limited pass band. It is therefore important to 'match' the various components of a system to obtain efficiency while maintaining the desired field response.

The hydrophone array used for most of this survey utilized Hall-Sears MP-4 hydrophones with a wide band pass. The seismic amplifier however, a Geospace 111-115, had a pass band from 10 to 300 Hertz (3 db. points). Because of

low frequency noise and the limited high frequency range, the filters were usually set to pass a band of frequencies between 80 and 300 Hertz.

After filtering, the signal is half-wave rectified before recording. This permits a phase correlation on the record, giving a clearer readout of reflected events. It is because of rectification that lines of alternate dark and light intensity occur on the record. Without rectification, only variation in tone would signify an echo sequence. The half-wave rectified readout is much superior to the full-wave tone correlation.

Resolution of Reflecting Horizons

Side by side recording of each shot and its subsequent echoes produces a profile of the sea bottom and sub-bottom structures. The width of the outgoing sound wave, however, tends to reduce resolution on the cross-section so produced.

The discharge of stored energy takes place over a finite time and the outgoing acoustical pulse is therefore stretched over a period of many milliseconds. The pulse is wavelike; that is, it consists of a series of compressions and rarefactions. After rectification in the recorder, and side by side recording, the result is a series of lines printed on the record corresponding to the compression and

rarefaction intervals arriving at the hydrophones. The first series of lines occurring at the top on the accompanying record plates is due to the wave travelling directly from transducer to hydrophones, its surface reflection, and the sound waves created by bubble oscillations and their surface reflections. The time and relationship at which these signals appear depends upon the distance between transducer and hydrophones and their depth below the water surface. The complete wave train appearing first on the record is called the direct arrival.

Following this down the record, the sea floor is represented by a similar band of parallel lines, for the echo is a close approximation to the output. Sub-bottom reflectors are likewise recorded by bands of lines rather than one discreet and sharp interface. The first line of any series corresponds to the true two-way travel time to the reflector. The other lines merely obscure echoes from closely spaced reflectors. Thus the resolution expected from the system is affected to a large extent by the length of the outgoing pulse. This is apparent in, for example, Plate XXIX between positions D and E, where sub-bottom reflections can be followed toward the sediment-water interface in Trincomali Trough. The actual outcrop of the reflectors is not apparent on the record because the sea floor reflection effectively

obliterates the sub-bottom for several milliseconds below the sediment surface.

The pulse length can be influenced to a degree by the towing depth of the transducer and hydrophones. The initial outgoing acoustical pulse is omni-directional, but upward directed energy is reflected from the water surface and re-directed downward, following the direct pulse in that direction at a time determined by the towing depth. The net effect is a stretching of the pulse time if towing depth is increased. The obvious correction is to tow the transducer and hydrophones near the surface. However, since penetration of sound is proportional to the inverse of the frequency (Grant and West, 1965), low frequencies, those favoured by deep towing, are desirable for maximum penetration. Ray theory indicates that by towing at a depth of one quarter of the dominant wavelength desired, interference generated by the direct and surface reflected waves should be constructive, increasing the outgoing energy at that wavelength. This assumes that the inital output pulse length is of sufficient duration for interference to take place. By towing at a depth of twelve feet, for instance, constructive interference should occur for a frequency of 100 Hertz. The time required for sound waves to travel to the surface and back to the transducer at that depth is five milliseconds. The discharge time of stored energy through the transducer is usually

greater than this, therefore interference will occur at that frequency. However, if the towing depth is increased much beyond this, unnecessary pulse stretching may occur. In actual practice, the sea surface is rarely a smooth reflector so that the depth to the transducer and hydrophones may vary considerably between wave crest and wave trough.

For this investigation, towing depth for both hydrophones and transducer was approximately 3 meters (10 feet). The dominant output frequency was approximately 120 Hertz. The pulse length for 5,000 joules of output energy as measured at the hydrophones for the direct arrival, was approximately 40 milliseconds. This is not compatible with high resolution and the resolution was not expected to be much better than 30 meters (100 feet). However, where the second reflector was particularly strong, it proved possible to resolve reflectors spaced at about 15 meters (50 feet) or in some cases less.

Reverberations, Multiples and Interference

Sound waves are reflected very readily from the sea surface due to the large acoustic mismatch between water and air. Thus energy reflected upward from the sea bottom is not only picked up by the hydrophones but is also reflected back again from the water surface. The reflected energy may be returned again and again from the sea floor to be picked up and recorded at each pass. The result on the record is a

series of irrelevant but related events which blanket weaker, but desirable, reflections from deep sub-bottom horizons. The effect is called sea bottom reverberation (Backus, 1959). Each reflection of the sea bottom on the record is a multiple of that event. In an area where reflection coefficients are large, many multiples may be present. Three or four bottom multiples are common and up to seven or eight have been observed, although this is by no means a limit to their number. Reverberation will continue until the energy is dispersed or attenuated. Plate XXX illustrates a record on which three multiples of the sea floor are present.

Reverberation is often stronger than sub-bottom reflections and tends to obscure these. Desirable information may therefore be lost on the record. In general, information is reliable down to at least the first multiple of the sea bottom. Beyond this, information may be garbled and unrecognizable unless from a strong reflector. Much of the ingenuity of signal processing methods devised to date is related to unscrambling signals from multiples, or repressing the multiples while enhancing meaningful reflections.

On occasion multiples may occur between pairs of reflecting sub-bottom surfaces, as when a strong reflector lies under the sea floor. Internal reflections may then take place between sea bottom and sub-bottom reflector. Sub-bottom reflected energy may also become trapped between

sea surface and sea floor or between a pair of sub-bottom reflectors. While these events are rarer than a sea floor multiple, they do occur and one should be aware of the possibility of their presence on a record. One of the problems of interpretation is the identification of true reflection horizons from those of multiply reflected signals. An example of a multiple reflection between sea surface and a sub-bottom horizons is illustrated in Plate X, positions E to F. A sub-bottom horizon at a depth near 0.275 seconds is repeated at a depth near 0.55 seconds after reflecting off itself as well as off the sea floor.

Effects of Topographic Roughness and Point Reflectors

The graphical record is affected by the geometry and structure of the sea bottom and sub-bottom. Because the transducer output is omni-directional, echoes will be received from objects on the sea floor ahead or beside the ship. Since the echo from the object has travelled a longer path than the echo from the sea bottom under the ship, it will appear on the record at a later time, that is, below the sea floor. As the ship approaches and passes the point, the echo path decreases. The recorded signal is thus seen to approach and perhaps pass through the sea bottom echo. If directly over the object, the recorded echo will peak above the sea floor then drop off below as the ship moves away. The resulting

recorded profile is a hyperbola projecting above the sea floor. Such hyperbolic waveforms emanating from point-reflectors or sharply terminating structures are often mistermed 'seismic diffraction patterns' and are common on CSP records where topography is rough and irregular, where steep slopes exist near the ship's track, or where sources internal to the sediments are present, such as boulders in till, or small gravel lenses in sand or silt. Examples of the waveforms from rugged topography are seen in Plate VIII, near position G.

Hyperbolae become asymptotic to 45° at horizontal distances that are large compared to depth to the reflecting source. At lesser distances they assume smaller angles, approaching zero degrees when the hydrophones are directly overhead. When such patterns are prolific on the record, it becomes difficult to separate true dipping strata from segments of these hyperbolae.

Even if the ship does not pass directly over a pinnacle or boulder but to one side of it, a hyperbolic pattern may still appear on the record but it may or may not break the surface of the sea floor. That is, a pattern may appear on the record which is not related to any reflector immediately under the ship's track. Such a reflection is termed a 'side-echo'. A reflecting source below the sediment surface which is under the ship's track may also produce the

same type of pattern. Bennett and Savin (1963) used two channel recording to distinguish side-echoes from true sub-bottom reflections. The true reflectors showed stronger records on a low frequency channel when compared to the high frequency recording.

Hyperbolae caused by surface irregularities can be a useful criterion to identify a particular reflecting surface if roughness is characteristic of that surface. However, such patterns obscure the true shape of an object, slope, or surface and thereby place a limit on the ability of sonic equipment to define shapes.

Apparent and True Slopes

As in echo-sounding, slopes recorded on a CSP profile are not true slopes. Aside from the fact that the survey may not have followed the fall line, horizontal and vertical scales are not usually the same. The vertical time scale is chosen from those available on the graphic recorder to suit the depth of water, sub-bottom penetration and the firing interval. The spatial equivalent of the vertical time scale varies with the velocity of sound in the materials traversed by the sound wave. The horizontal scale depends upon the paper speed and speed of the ship over the bottom. Most commonly, the vertical scale is greater than the horizontal, causing small angles to be exaggerated. Because

transducer and hydrophones are omni-directional, or semi-directional, side-echoes can be received from points on the sea floor (and sub-bottom) which are not directly under the ship. As a slope is approached, echoes are received from a point tangent to the wave-front before the ship actually arrives over it. This point is therefore at a shallower depth than the sea floor under the ship. On the record then, the slope appears to be less than its actual value. The real and apparent recorded slopes are related by the trigonometric function $\sin \theta = \tan \phi$ where θ = angle of the true slope and ϕ = angle of the recorded slope (Krause, 1962). The error for slopes up to 15° is small, less than 3%. For slopes greater than this, the error may become appreciable.

Sub-bottom slopes can be corrected in the same way but the sound velocity in the overlaying sediments must be known accurately, especially if one end of a slope is buried deeper in overburden than the other. In such a case the time differential due to depth variation in the sediments can be important in determining the real angle.

Nature of Seismic Reflectors

The usual objective of a continuous seismic survey is to investigate the sedimentary or rock strata in the upper layers of the earth's crust. How reflections as recorded on the seismic profiles are related to geological strata is a

question fundamental to the principle of seismic methods. Knowledge of the answer assists a confident interpretation of the seismic profiles.

It is a common misconception that acoustic reflections occur only at seismic velocity discontinuities. According to wave theory, (see, for example, Officer, 1958; Lindsay, 1960) acoustic reflections occur at a change of acoustic impedance, a quantity defined as the product of density, ρ, and acoustic velocity, V. Thus, while velocity discontinuities without a corresponding change in density will cause reflections, density contrasts with no velocity change may also give rise to reflections. Grain size, mineral composition, porosity, degree of lithification, depth of burial, as well as other factors, affect the elastic properties of materials upon which the acoustic impedance depends. Velocity is especially sensitive to changes in material properties. One has only to glance at a velocity log of a well to see how rapidly this varies in a natural situation. Density and velocity may also change rapidly in the lateral direction. Because of the many geological factors affecting both parameters (Nafe and Drake, 1957; Hamilton, 1959) the variation of acoustic impedance can be, and is, a complicated function. However, because the acoustic impedance is intimately related to the geology, the seismic reflection method does work well.

Identification of reflectors as specific rock or sediment types cannot be made on the basis of seismic evidence alone. Additional geological control in the form of borehole logs, sea floor samples in areas of outcrop (ascertained from the seismic record), and study of local and regional land geology is necessary to make more-or-less positive identification possible. However, from past experience in seismic profiling, the general geological nature of the material traversed may be surmised from the record character itself. Moore (1960) states that the strongest sub-bottom reflections occur from the sea floor and from bedrock surfaces beneath relatively soft sediments. In some cases they occur from bedding planes in contrastingly layered sediments, such as gravel or sand lenses in clay, etc. Moore also found that silty-sand overburdens show acoustic homogeneity. He interpreted reflectors within a sedimentary layer, all parallel to the profile of the sea floor, as an indication that the present sea floor is a depositional environment. The reflectors were usually from a material of contrasting texture to the sediment mass, i.e. a sandy layer in silty clay or gravel in sand. McLure, et al, (1958) found that the reflection record was layered where the sediments contained thin sand or silt beds and thin shell lenses. One thin sand bed only 0.4 wavelength thick produced a prominent reflection over a wide area in which it was present. In

other areas, reflections which did not correlate with lithological composition were shown by laboratory analysis to have a probable correlation to changes in moisture content of the sediments. Beckman, et al, (1959) showed a dependable correlation between the continuous seismic reflection record and changes in porosity. The change in number of hammer blows required to drive a sampler a unit depth into the sedimentary column also correlated with depth to reflecting surfaces.

From practical application then, as well as from theory, strong sub-bottom reflections should be generated from such interfaces as occur between unconsolidated and semi-consolidated sediments, or between indurated or compacted bedrock and unconsolidated and semi-consolidated overburden. Where such major changes in the nature of the material occur, reflections should be recorded unless the reflecting surface lies so deeply buried that reflected energy is too greatly attenuated before reaching the hydrophones. Internal reflectors within an otherwise uniform layer may represent factors other than lithological changes but should, nevertheless, be indicative of internal stratification of the sediments. Seismic reflections on the records should therefore show evidence of such features as bedding planes, truncated layering, faults, synclinal and anticlinal structures, unconformities and disconformities; in short, normal and expectable geological structure. Following Curray and Moore (1964,

p. 202), such structure can be interpreted from 1) lateral continuity of reflectors, 2) attitude of reflectors with respect to the underlying, overlying and neighbouring strata, 3) the nature of the upper contacts of sequences of reflectors, and 4) the thickness of the units, both individually and in sequence.

Thus, as Curray and Moore (op. cit.) wrote, reflections arise from "positions of the former sea floor which existed as interfaces for a sufficiently long time (prior to burial) to develop into acoustic reflecting horizons by changes in lithology, by dessication under sub-aerial conditions or by more complete consolidation. To develop such characteristics the fossil surfaces must have formed during changes in the depositional, oceanographic or sea level conditions."

CHAPTER III

MORPHOLOGICAL SUBDIVISION OF THE STRAIT OF GEORGIA STUDY AREA

As is often the case in oceanographic studies, no topographic chart of the study area was available. Therefore a contoured base chart, exhibited as Figure 5, and found in the back pocket of this thesis, was prepared by the author from the following data: 1) hydrographic charts of the area, 2) field sheets of the Canadian Hydrographic Service, and 3) echograms obtained by Cockbain (1962 a). As with the Hydrographic Service, the echograms were plotted assuming a sound speed in water of 1,465 meters per second (4,800 feet per second). The echograms obtained by Cockbain covered only the northern part of the present study area from Sand Heads to Ballenas Islands. In the southern part of the area, from Sand Heads to Patos Island, no echograms were obtained. However, more Hydrographic Service field data sheets were available. The bathymetry of this area was drawn mainly on the basis of these field sheets. As the topography tends to be smoother in the southern region, no important detail that could be shown in the 10-fathom (1 fathom = 1.83 meters or 6 feet) contour interval was considered to be lost. Where information density was low, as at Gabriola Reefs, contours were omitted. On the basis of having later

returned directly to charted points, the overall accuracy of the positions of contours is considered to be better than 0.5 kilometers.

I. TERMINOLOGY OF SHELF- AND SLOPE-LIKE FEATURES

Water depths in the study area reach a maximum of 433 meters (237 fathoms), well beyond the depths normally found on continental shelves. In fact, more than 50% of the area involved lies at depths greater than 132 meters (seventy-two fathoms), the world average for the depth of the shelf edge (Shepard, 1963, p. 257). We are therefore dealing with a body of water deeper than continental shelves but not as deep as ocean basins. Table I lists the area and percentage area in various depth zones.

TABLE I

DEPTH ZONES IN THE STRAIT OF GEORGIA

	Square Nautical Miles	% of Total Area
Total Study Area	1,100	100
Area Deeper than 20 fathoms	823	75
Area Deeper than 100 fathoms	477	43.5
Area Deeper than 200 fathoms	106.6	10.3

Because the area is shallow compared to deep ocean basins, but deep compared to continental terraces, the usual meaning attached to the nomenclature of the marginal areas of the continents is not valid here. For instance, in many cases water depths very near shore may approach or exceed the depths normally associated with outer continental shelf or slope areas. These terms are therefore redefined here in the special context of their use in this thesis. For this purpose a shelf, where it exists, is defined as the area adjacent to the shoreline which extends seaward at a shallow declivity. In most of the study area the shelf is not wider than two to three kilometers. In some areas, no shelf exists. The shelf edge occurs at any depth and is defined as the point where the declivity increases markedly. The slope, for the purpose of this thesis, is the inclined area which connects the shelf to the basin floors. In areas without a shelf, the slope begins at the shoreline. Where it is necessary to make reference to a particular shelf or slope on one or the other side of the Strait, the words 'mainland' or 'island' will be prefixed, where 'mainland' refers to the region on the north and east sides of the Strait, including any islands such as Bowen Island, and 'island' refers to the south and west side of the Strait including Vancouver Island and the Gulf Islands. Any further distinction will be indicated by the place name. At the Fraser Delta, the terms 'shelf' and 'slope' are not

directly applicable. Instead the terminology of Mathews and Shepard (1962) will be followed. Here, the upper slope of the delta front constitutes the area of steepest decline, averaging from $1\frac{1}{2}^{\circ}$ to $3\frac{1}{2}^{\circ}$. At greater depths, the declivity is markedly less. These are the lower slopes of the delta. The upper and lower slopes are composed of foreset beds. While Mathews and Shepard did not recognize bottomset beds, in this thesis they are found to occur at some distance from the river mouth as basin fill.

II. SUBDIVISIONS OF THE STUDY AREA

To simplify the presentation of data, it is convenient to apply names, not only to individual topographical features, but to those areas of submarine topography which, from their bathymetry or unique character of geological or geophysical data, may be grouped as a unit under one descriptive term or heading. Previous classifications have been made only in local areas or under more general groupings. For instance, Waldichuk (1953) divided the Strait into three areas on the basis of sediment distribution. The southern section in the region of the Gulf Islands and San Juan Archipelego, including the bays and channels, were described as an area of diverse bottom type with little deposition and possible bedrock erosion. The central Strait, covering much of the area of this thesis, was recognized as dominated by soft muds

deposited as a result of Fraser River inflow. He distinguished the Strait north of the present study area by its sand, thought to be derived from erosion of shore sediments in that region. While this classification is valid as a regional description of facies, it bears little relation to the morphological features of the sea floor, or to their geophysical character.

Cockbain (1963 a) subdivided the northern part of the study area into three morphologically distinct divisions: the Fraser River Delta in the southeast; an area of banks and subdued topography along the mainland side of the Strait in the northeast; and an area of deep basins separated by ridges along the Vancouver Island coast. Although this classification is limited, it can be enlarged, defined more clearly, and made of use here.

On the basis of topography alone, the study area is easily divisible into a northwestern region of rugged character, a central smoother region dominated by the Fraser Delta, and a southern area where the Fraser River has not deposited recent sediments and erosion may be occurring. Topography of the southern area is intermediate in ruggedness, being more irregular than the area subdued by a thick cover of river-supplied sediment but not as irregular as the northern region of banks and basins. However, CSP profiles indicate that a further subdivision, based on more than topographical expres-

sion, is useful. The character of the seismic record is a function, although not a unique function, of the geology of the bedrock and soft sediments through which the seismic sound waves pass. Thus a change in character of the seismic record may be related to a change in character of the rocks or sediments traversed. In the Strait of Georgia the record character is consistent over certain areas. These areas are usually related to the bottom configuration, indicating some consistency in geology as well as topography. The following subdivision of the Strait is based on morphology, bathymetry, geological structure and geophysical character. The subdivisions are shown on Figure 5.

Subdivision of the Study Area

1. Fraser Delta Area
2. Area of Deep Basins
3. Elevated Area of Ridges
4. Roberts Swell and the Nearby Mainland Shelf
5. Boundary Basin and Alden Ridge
6. Island Slope

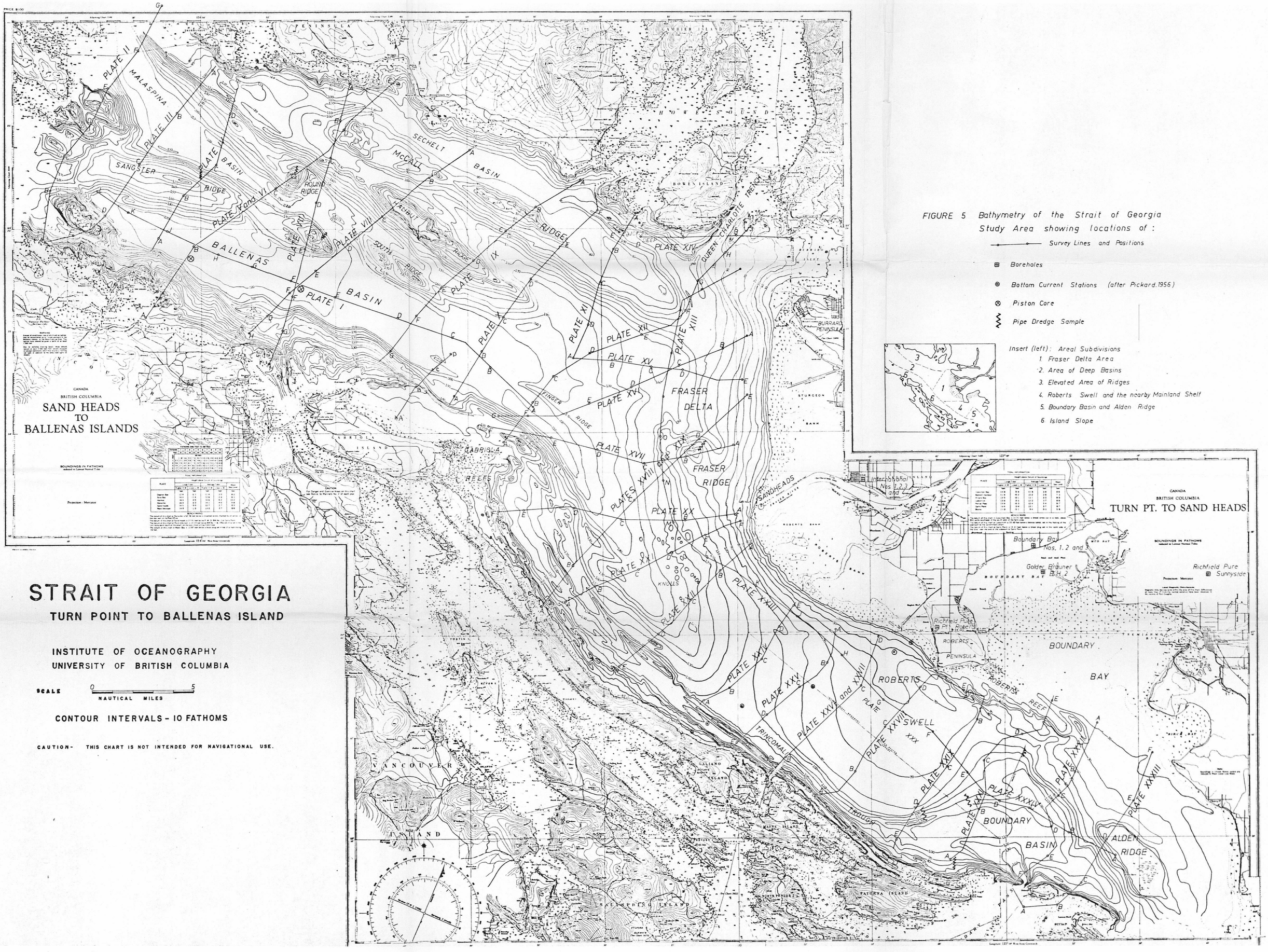


FIGURE 5 Bathymetry of the Strait of Georgia Study Area showing locations of :

- Survey Lines and Positions
- Boreholes
- Bottom Current Stations (after Pickard, 1956)
- Piston Core
- Pipe Dredge Sample

Insert (left): Areal Subdivisions

1. Fraser Delta Area
2. Area of Deep Basins
3. Elevated Area of Ridges
4. Roberts Swell and the nearby Mainland Shelf
5. Boundary Basin and Alden Ridge
6. Island Slope

CHAPTER IV

SEISMIC DATA

The seismic data gathered during this survey is presented in Plates I to XXXIV. Photographs of the original records are given, along with line drawings showing the geophysical and geological interpretation of the profiles. These should be referred to in relation to the text.

Tracks of the survey lines are indicated on Figure 5. Positions of fixes along each line are marked by circles and assigned a letter. On each profile of Plates I to XXXIV the equivalent position is marked with the same letter. On the photographs, the times of the original fixes, recorded by vertical lines, can be seen on the records below the letters.

Difficulties in presentation of this CSP data are caused by the large quantities of information, the diversity of geology and the desire to be concise, consistent and logical. Twenty-six profiles, representing 600 kilometers of line, cross the Strait in a transverse direction. In addition 126 kilometers of line follow the axis of the Strait. Shorter profiles in other directions make a total of 790 kilometers of line whose information is to be presented. Other authors have used various techniques of data presentation. However, rather than each profile being presented and explained in the

text, the data in this study will be presented by area, the areas being those listed previously under the heading "Morphological Subdivision of the Strait of Georgia" and indicated on Figure 5. To make clearer the seismic data, major horizons have been outlined in heavy line on the line drawings. These represent units separated by major unconformities corresponding to geologically significant time breaks. Thinner lines represent layers which can be distinguished and separated within units. The thinnest lines show the shape and attitude of the internal bedding or stratification within a layer, without intending to give special meaning to any particular reflector. Rather it is the trend of the internal stratification which is significant in most cases.

The geological interpretations of the profiles are made on the line drawings. A legend opposite Plate I identifies the symbols used.

Because of the difference of vertical and horizontal scale, angles on the profiles are greatly distorted. Low angles have the appearance of steep slopes. Exaggeration is useful in detecting and mapping gentle slopes or small changes in low-angled slopes, however, steeper slope angles are compressed, making it difficult to measure or distinguish one angle from those around it with similar, but not equal, slopes. The angle at which exaggeration changes to compression depends upon the inverse of the exaggeration factor. For an

exaggeration factor of 12, that most common on the profiles of this thesis, angles less than 16 degrees are made larger, while those greater than 16 degrees, while still exaggerated, are compressed or squeezed together. Thus it becomes virtually impossible to tell from an exaggerated record if a steep slope is, for example, 60 degrees or 70 degrees. Nomographs on the line drawings show the exaggeration.

One difficulty encountered in this thesis was in separating true strata dipping at angles or 30 degrees or greater, from hyperbolae caused by point reflectors. The hyperbolae are asymptotic to 45 degrees when the reflecting point source lies under the ship's track. Various techniques were used to distinguish between these features. For instance, if carried far enough, the real and hyperbolic patterns, if not at the same angle, must converge and cross. However, most bedding reflectors do not extend far enough for this to occur. True bedding planes may change angle, usually becoming less steep with depth. This was a useful parameter but one which did not always occur. When no other method of distinguishing bedding presented itself, hyperbolic patterns were drawn to scale complete with the local exaggeration and used as templates to try to determine the source of the reflection.

The vertical exaggeration varies from record to record and, in fact, probably varies over the length of any one record. Variation was caused mainly by changes in ship's

speed over the ground which was in turn caused by variations in speed and direction of tidal currents encountered in the Strait. The effects of wind, too, which at times during the survey reached gale force, also contributed. In one instance the difference between course steered and course made good amounted to 30 degrees as the vessel made leeway in high wind and heavy tide. Even should perfect control of ship's speed be obtained, exaggeration would still vary because of variations in the velocity of sound through different sediments. The nomograph provided on the line drawings of the profiles should therefore be used to measure sea floor slopes only. Measurement of sub-bottom slopes requires knowledge of sediment velocities. These are not well known, but sub-bottom slopes and sediment thicknesses have been estimated using velocity information from available sources. Velocities in various sediments are discussed in the Appendix. The vertical exaggeration in the profiles, calculated for a sound velocity of 1,460 meters (4,800 feet) per second, ranges between nine and fifteen, with twelve the most common. One profile (Plate VIII) recorded at a paper speed slower than normal, has an exaggeration of slightly more than four. Records obtained from Shell Canada Limited have exaggerations of seven to ten.

I. THE FRASER DELTA AREA

In this thesis, the name 'Fraser Delta' will refer to the submarine portion only of the Fraser River Delta.

The region termed here the Fraser Delta Area is that area which is, in general, most heavily encroached upon by the advancing sediment front deposited in the Strait by the Fraser River. It is somewhat arbitarily chosen to represent the delta region but it is not meant to infer that sedimentation from the Fraser River is negligible outside of this area. In fact, sediments from this source occur throughout much of the study area. However, the area from Point Grey at Vancouver seaward seven kilometers to the west, then southwest to the deepest point in that direction, following the axis of the deepest part south to the 49th parallel, then in a direct line to Point Roberts, represents the main portion of the foreset beds of the delta where deposition is heaviest.

The Fraser Delta was described by Johnston (1921) and later by Mathews and Shepard (1962). Its active part extends into the Strait on a broad front of more than thirty-seven kilometers from Point Grey to Point Roberts. Extensive tidal flats, as much as nine kilometers wide in places, extend from the sub-aerial delta and dry out at extreme low tides. These flats, called Sturgeon Bank north of the main

river mouth at Sand Heads, and Roberts Bank to the south, are terminated seaward by an upper slope averaging 1 3/4 degrees to $3\frac{1}{2}$ degrees and locally steeper (Mathews and Shepard, 1962). This upper slope tends to decrease with depth, merging with a lower, more gentle slope which continues downward to the adjacent basin depths.

From Point Grey to Sand Heads, the general trend of the delta front is north-south, in marked contrast to other features in the Strait which trend in the general northwest-southeast direction of the Strait itself. South of Sand Heads the delta front sweeps to the southeast and, below 100 meters (55 fathoms), merges with the contours of Roberts Swell.

Northwest of Sand Heads the only major disruption to the generally smooth contour of the delta front is a protruding ridge top noted by Mathews and Shepard and thought by them to possess a core of Tertiary or Cretaceous bedrock. The peak of this ridge, named here Fraser Ridge, reaches to less than 145 meters (80 fathoms) of the water surface. It is almost buried by the advancing delta.

West of Sand Heads a group of small hills with relief of about fifteen to thirty meters occurs beyond a depth of 220 meters (120 fathoms). These hills or knolls, located by Mathews and Shepard, are distinctly a feature of deltaic or estuarine sedimentation and, as shown by the continuous

seismic profiles over the area, are not an expression of buried topography. Mathews and Shepard suspect the hills to be remnants of former landslides, modified by current action and later sediment.

The sub-bottom configuration of the delta front is illustrated in eight profiles which cross the Strait in a transverse direction and one profile along the front of the delta. Two of these profiles were recorded by Shell Canada Limited and represent identical track lines: one profile recorded the output of a sparker source (Plate XVIII); the second (Plate XIX) was obtained with a high-powered gas exploder source used contemporaneously with the sparker source. Photographic reproductions of the original records are shown with line drawing interpretations in Plates XV to XXIII. The ridges evident on the delta profiles will be discussed in a later section. Similarly the bedrock reflector beneath the delta sediments will be discussed with the section on the island slope.

Of the slopes measured on the profiles of the delta, none, except at minor local disturbances, exceeds an apparent angle of $2\frac{1}{4}^{\circ}$. Upper delta slopes are steepest. Below 220 meters (120 fathoms) the slope angle is reduced to about 1°, and decreases further as water depth increases.

Sediments comprising the modern delta front form a thick wedge extending at least fifteen kilometers across the

Strait from Sand Heads to the opposing island slope. Since Sand Heads is itself eight kilometers from the shore line of the subaerial delta the thick marine delta sediments extend at least twenty-three kilometers into the Strait. Under the thick eastern part of the wedge, the base of the sediments cannot be seen on the profiles. However, as the wedge thins to the west, the pre-delta surface becomes apparent rising toward Vancouver Island. On this side many high rocky pinnacles have been buried, or nearly so, by the advancing delta sediments, some two million cubic meters (7×10^8 cubic feet) of which are added to the delta front each year by the Fraser River (Mathews and Shepard, 1962, p. 1424). These buried rocky pinnacles form part of a ridge system paralleling the Gulf Island slope south of Gabriola Reefs.

Under the sediment-water interface, the character of the CSP records may be related to the composition of the sediments. Similarity in the character of echograms was used by King (1965) to delineate areas of like sedimentary facies on the sea floor off Nova Scotia. Profiles over the Fraser Delta show changes in character from the steeper upper slopes, above the 185 meters (100 fathoms) contour, to the flatter lower slopes. The steeper upper slopes are characterized mainly by low seismic penetration and a pattern of random sub-bottom echoes. Lower delta slopes everywhere permit

greater penetration and show more organized and stratified patterns in the sub-bottom reflections.

The proportion of unconsolidated sand in the sediment has the inverse characteristic to the seismic penetration (Mathews and Shepard, 1962, Figure 2). On the upper slopes sand content is high, especially south of the river mouth. On the lower slopes sand content decreases while silt and clay fractions assume greater proportions. In other areas where seismic penetration is very good, as in Ballenas Basin, sampling has shown the sea floor is composed almost completely of silt and clay sized particles. It is only when the upper delta front is close that signal losses become excessive. This points to a conclusion that loose unconsolidated sands of the delta front act to reduce penetration of sound waves. Other workers have also found that clean sand prevents seismic penetration and obscures the underlying section (Moore, 1960, p. 1124; McLure, et al, 1958; Ostericher, 1965, p. 42). Other characteristics of the sediment as well as sand content may have the same effect, but if so, these are not known at the present time.

According to Johnston (1921), no stratification of sediments is visible in cores from water depths of 90 to 180 meters (50 to 100 fathoms), but massive bedding may be present. He considered the fore-set beds of the delta are laminated whereas bottom-set bedding lacks lamination. In Johnston's

terminology, bottom-set beds occurred at depths greater than 90 meters (50 fathoms). Mathews and Shepard (1962, p. 1427) found only slight evidence of lamination in cores taken mainly from the deeper slopes. They disagreed with Johnston's terminology and preferred to call all bedding in the area of their investigation 'fore-set' beds more in keeping with the slope exhibited by them. The CSP profiles across the delta area show that sediments at the present day water-sediment interface would all come under the term 'fore-set' whereas much deeper below the fore-set beds, older flat-lying strata are probably bottom-set bedding. The modern delta sediments are symbolized by H_3 on the line drawing interpretations while the older buried bottom-set beds are given the symbol H_1. Modern flat-lying bottom-set beds also occur in the deep basin area and are marked H_2.

The seismic profiles over the delta, although not expected to resolve thin laminae, give little indication of stratification in the upper delta slopes. The presence of cross-lamination or other small-scale stratification that cannot be resolved seismically is not precluded, but large-scale continuous bedding such as is present over other parts of the study area cannot be seen here. As water depths increase, stratification also increases until, under the lower delta slopes, many sub-bottom reflectors are observed to lie conformably, or nearly so, below the sediment-water

interface. However, stratification does not continue throughout the entire column of delta sediments but ceases as depth of overburden increases. Stronger seismic reflections at lower depths can still be seen marking major horizons and sediment-bedrock contacts. The fact that the upper reflections are weak, and do not occur beyond 90 to 120 meters (300 to 400 feet) below the sediment surface, implies that the sediments are nearly homogeneous and differences in acoustic impedance are not great. The differences, although possibly caused by variations in parameters such as density or porosity (Hamilton, 1959; McLure, et al, 1958), are more likely associated with depositional bedding, perhaps located below sampling depth. Furthering this impression is the fact that where ridges have interrupted the smooth transport of sediments, the stratification is also disrupted but tends to follow an expectable pattern of depositional bedding, rising over ridge tops and down the other side. In fact, reflectors in sediments below the ridge tops end abruptly against the sides of the ridge, indicating the ridges ponded sediments until the ridge was completely buried, at which time the sediments moved over the top. Eventually, external evidence of the buried ridges was destroyed, leaving in some cases no evidence at all, in others only a hump or small scarp remaining on the sediment surface to mark the burial.

In the area of anomalous hummocks, or knolls, off the Fraser River mouth, stratification in the H_3 layer is severely disturbed, but not in a random manner. Instead it is bent up under the knolls, and down between them. Except where the knolls have been buried by younger sediments, the stratification follows the surface topography. Later sedimentation has not buried all the hills evenly, but tends to bury those hills on the higher slope first. This is highly suggestive of burial through a mechanism of bottom transport of material from a higher delta source rather than through settlement of material in suspended transport from the water above. The anomalous knolls will be treated in a later section.

Much deeper under the delta sediments, every profile records a strong reflection from an almost flat-lying horizon at a depth below sea level of nearly 0.6 seconds travel time. This horizon marks the top of the H_1 layer. Below the horizon, secondary reflectors in the H_1 layer indicate that the layer is stratified. The flatness of the H_1 stratification throughout the delta area contrasts with the dipping strata in the overlying H_3 layer. The H_1 layer can be correlated from record to record under the whole of the western part of the delta where it is confined by the basin sides. At Roberts Swell in the south, it disappears, probably by wedging out over the Roberts Swell sediments. However, no information

is available at the present time to indicate the exact relationship between these two sedimentary units. Between the adjacent profiles of Plates XXII and XXIV (see Figure 5 for the locations of these profiles), the H_1 layer disappears.

Under Ballenas Basin to the northwest (Plate I) a similar correlatable horizon extends the layer as far as the end of the basin in that direction. Plate XVI shows the layer also exists at a slightly higher elevation of 0.55 seconds between Point Grey and the buried extention of McCall Ridge. From there it can be traced farther to the north deep under the sediments of Queen Charlotte Trench (Plates XII and XIII), but wedges out against a rising bedrock sill (Plate XIV) before entering Howe Sound. It is not known if it continues on the other side of the sill in Howe Sound.

The H_1 horizon dips to the northwest at an average of no more than one part in one thousand, or about four minutes of arc, beneath Ballenas Basin. A slight cross-basin tilt toward Vancouver Island in the region of the Fraser Delta is probable but may reflect a variation in speed of sound through the overlaying wedge of modern delta sediments. Reflectors internal to the H_1 layer show little turn-up on the flanks of the confining bedrock sides. Sediment transport was therefore mainly along the length of the basin, rather than transversely across it. In fact, from the slope of the reflectors, transport was almost certainly from southeast to northwest,

away from the Fraser River region. These sediments are therefore related to the Fraser River drainage system and are probably ancient bottom-set bedding of the Fraser Delta. Because the H_1 horizon is well below Fraser Ridge, even in its buried parts, and because the cross-basin slope is so slight, at the time of deposition the sub-aerial delta must have been some distance from its present site, perhaps many kilometers to the east or southeast of Fraser Ridge. Near Point Grey, and in Queen Charlotte Trench on the east side of Fraser Ridge, the H_1 horizon lies about 8 to 15 meters (25 to 50 feet) above the same horizon west of the ridge. From this it can be inferred that Fraser Ridge and McCall Ridge are connected near Point Grey and so blocked the former movement of sediments into Ballenas Basin around the north end of Fraser Ridge. A break in this ridge system which allowed sediment transport into the basin may be buried somewhere to the south of the present river mouth.

The thickness of the H_1 bottom-set layer is variable, depending upon the relief of the underlying bedrock. Under the fore-set beds of the delta it averages about 80 meters (260 feet) and sometimes exceeds 150 meters (500 feet). Deposition occurred directly over bedrock and semi-consolidated deposits forming Ballenas Basin floor. In the depositional process, irregularities of the floor were filled in, leaving a relatively flat, smooth surface over which modern fore-set

bedding has advanced. The ancient bottom-set beds were contained by the walls of the basin in much the same manner as is presently occurring in the northwestern basin area.

East of Fraser Ridge the top of the bottom-set bedding occurs at about 400 meters (1,300 feet) below present sea level. If the same elevation persists to the sub-aerial delta, bottom-set bedding would only be found in wells deeper than 400 meters. Few wells anywhere in the delta penetrate that deeply. Those that do have been too poorly logged to differentiate bedding types.

Total thickness or volume of all delta sediments cannot be directly calculated because the pre-delta surface is not detected under the thickest section of the delta. West of Fraser Ridge and Sand Heads, sediments reach a maximum measurable thickness of at least 290 meters (950 feet). At a position about 3.5 kilometers southwest of Point Grey, measurable sediment thickness on the upper slope is approximately 275 meters (900 feet). The indication there is that the thickness increases much more to the east. With bottom-set bedding at 400 meters, sediments should be at least that thick in depressions of the sub-delta surface. A drill hole near Steveston at the mouth of the Fraser River is reported to have penetrated 215 meters (700 feet) of sand and recent delta material before striking a large boulder (Johnston, 1919, p. 6) but it is apparent from Plates XVII, XVIII and XIX, that the

pre-delta topography had considerable relief. These profiles, which outline the buried basin under the delta sediments indicate Fraser Ridge at one time stood almost 370 meters (1,200 feet) above the pre-delta basin floor. Most of the relief of the pre-delta surface has been completely buried. Only locally, such as at Fraser Ridge, does any evidence of it exist on the present sea floor.

II. THE NORTHWESTERN BASINS

In the northwestern section of the study area, two deep basins, Ballenas and Malaspina, occupy the western floor of the Strait.

Despite the great depth of water, good seismic penetration of up to 0.25 seconds or more was achieved in the unconsolidated sediments here. Nine profiles, shown in Plates III to XII, cross Ballenas Basin to the island slope. Plate II passes north of the end of the basin. In addition, Plate I records a profile along the length of Ballenas Basin from a point near Gabriola Reefs in the southeast, to the end of the basin north of Ballenas Islands. Plates II to VII also cross Malaspina Basin but no profile was obtained along its length.

The bedrock sides of the northwestern basins are more directly associated with either the island slope or the elevated ridge areas, and they will be discussed in those

sections. Similarly the bedrock floor under the sediments is also referred to the same sections.

Except for the southwestern part of Ballenas Basin, little penetration was achieved in the basement rock underlying the sediments. This is hardly surprising since even the western side of Ballenas Basin, which is bare of sediment, offers little penetration, especially northwest of Gabriola Reefs. Locally, some bedrock reflectors do occur under the basin sediments but, as there is little continuity to them, their significance is obscure.

Ballenas Basin

Ballenas Basin extends along the base of the island slope from Ballenas Islands in the northwest to Galiano Island in the south, a distance of more than sixty-five kilometers. Passing east of Gabriola Reefs, the basin turns to the south and is now terminated by the rising slope of the Fraser Delta southwest of Sand Heads. The flat floor of the basin is from four to six kilometers in width and for the most part exceeds 370 meters (200 fathoms) in depth. The sea floor slopes from the delta toward the northwest, merging smoothly with the floor of the basin in that direction. The slope, decreasing continuously from the delta, becomes almost horizontal in the deepest part of the basin. With the exception of the delta area, most of the basin perimeter is ringed by the relatively steep sidewalls of the island slope on one side and submerged

ridges on the other. At the delta front, fore-set beds have filled much of the basin from the side and the lower fore-set beds now rest well up the opposing island slope.

Near Gabriola Reefs the peak of a linear ridge extends from under the delta sediments, pointing northwest down the middle of the basin. This ridge will be called Finger Ridge. It now lies almost buried but has sheltered the section of basin on its southwest from the full effects of delta sediments. The basin in the lee of the ridge remains deeper than the side closer to the Fraser River.

Two major sedimentary layers can be distinguished in Ballenas Basin. In the upper layer, many strong internal reflectors dip to the northwest from the Fraser Delta. These sediments are therefore a part of the modern delta. In the deep basin area the slope decreases to near zero. The upper layer of sediments can therefore be called modern bottom-set beds of the delta. On the line drawing interpretations, these pro-delta sediments are marked H_2. The H_2 layer thins from the delta area to the northwest and some reflectors disappear in the basin (Plate I). Thus, as the distance from the delta increases, fewer reflectors occur in the sub-bottom sediments. Some, however, continue to the end of the basin. As explained later, at least some continuous reflectors may be due to turbidite layers in the basin.

Below the modern pro-delta sediments, a strong reflecting horizon marks the top of the H_1 layer of ancient pro-delta sediments. This layer is correlated with the H_1 layer under the delta area and is continuous to the end of Ballenas Basin. It covers and smooths the relief of the bedrock upon which it rests leaving in general a flat, uniform surface similar to the present sea floor of the basin. The similarity to modern day pro-delta sediments strengthens the assumption that this layer does represent an older episode of the delta.

The volume of sediments in Ballenas Basin can be estimated. The upper layer of H_2 sediments has an average depth of about 76 meters (250 feet). Since the basin sides are steep and the base of the layer is a reasonably plane surface, a polar planimeter was used to measure the area over which these sediments lie. Measuring to the 366 meter (200 fathom) contour at the delta yields an area of approximately 250 square kilometers. The volume of H_2 sediments contained in Ballenas Basin is therefore 12 cubic kilometers. If the delta sediments (H_3) are included, taking only the volume west of Fraser Ridge below 183 meters (100 fathoms) where the total sediment column can be seen, the volume of modern sediments totals approximately 60 cubic kilometers.

The volume of the H_1 layer is more difficult to determine, but a rough estimate can be made, considering the

shape of the bedrock floor and varying depth of sediment over it. The volume of sediment including the H_1 layer under the Fraser Delta area west of Fraser Ridge is approximately 12 cubic kilometers.

Malaspina Basin

North of Ballenas Basin, and separated from it by a high ridge, a second deep basin, Malaspina Basin, enters the study area from Malaspina Strait between the mainland and Texada Island. Only the most southerly part of this basin extends into the study area and no information other than that on nautical charts is as yet available for more than that part shown in Figure 5. However, this appears to be the deepest section of a long, narrow basin whose depths over a large area mainly exceed 370 meters (200 fathoms). At least one extensive deep occurs in the basin containing the greatest depth of the whole of the Strait of Georgia, 433 meters (237 fathoms). The deep area may be related in some way to a high, steep-sided, round knob rising from the side of the basin. The knob, called Round Ridge, constricts the basin to a narrow passageway at that point. The floor of the passageway is tilted toward the base of the knob with the deepest point occurring at the base.

Not far from Round Ridge, Malaspina Basin joins the central portion of Ballenas Basin. The floors of both basins are concordant at the confluence but differences are apparent

in the basins themselves (see Plate IV). Whereas Ballenas Basin is generally flat floored, the floor of Malaspina Basin may bow up or sag down in the centre. It tends to flatten in the north as it leaves the study area. Malaspina Basin is, despite its deep areas, slightly shallower on the average than Ballenas Basin.

Since only a few profiles cross Malaspina Basin and no axial profile was recorded, less is known of this basin than Ballenas Basin. Two sedimentary layers are found here. The upper layer is almost devoid of internal reflectors. A similar reflectionless layer occurs over the nearby elevated ridge area. From the absence of internal reflectors and the good transmission of sound through it, it has been called 'seismically transparent' and given the symbol H_t on the line drawings. Sediments in which no internal reflectors occur are considered to be homogeneous (Moore, 1960). This characteristic does not occur in the upper layer of Ballenas Basin, and therefore denotes a difference in the sedimentary history of the two basins. The sediments are undoubtedly Holocene in age, possibly hemipelagic, with the most probable source being the Fraser River, although the possibility of unknown sources to the north of the study area should not be overlooked. The sediment surface slopes to the southeast as opposed to the sediments in Ballenas Basin, but this may be due to an initial southeast slope on the lower layer.

Below the transparent or H_t layer, some strong reflectors occur in the deep sediments. Correlation with similar reflectors in Ballenas Basin is difficult because of the number of choices and the lack of an axial profile down the basin. The top of the H_1 layer of Ballenas Basin may be represented by the similar strong reflecting horizon in Malaspina Basin at the base of the H_t layer. Insufficient data prevents making a good correlation except, perhaps, near the junction of the two basins. Reflectors in the lower layer are not continuous throughout the section of basin studied.

Deep reflectors in both Malaspina and Ballenas Basins indicate a central sag, possibly due to sediment compaction with time and depth of burial, although other factors such as currents cannot be ruled out. The present sea floor is warped and therefore other forces may be operating to shape it.

Because of uncertainties in age, source and history of the deep sediments of Malaspina Basin, they are denoted by the symbol Q on the line drawings. The possibility exists that some deep sediments may be Late Pleistocene in age.

Sediment Thickness and Rate of Deposition

Total sediment thickness in the basins varies with relief on the bedrock below the fill. Maximum thickness may be as much as 260 meters (850 feet) in Ballenas Basin at points away from the delta front area, and less, about 230 meters (750 feet) in Malaspina Basin. The bedrock floor of

the latter basin is not known to be as deep as that of Ballenas Basin although if a profile was made along its axis, deeper areas might be found. Over the length of Ballenas Basin, the average thickness is approximately 200 meters (650 feet). Without an axial profile in Malaspina Basin, it is difficult to estimate an average there.

An average rate of sedimentation can be calculated for Ballenas Basin. However, many variables occur which would seem to make any figure unrepresentative of the true rate. Sediment supply must have varied considerably since Late Pleistocene as the more accessible glacial deposits were removed and slopes became more stable. Thus the present rate of sedimentation may be less than when the delta was first building. Counteracting this is a reduction in distance from the river mouth as the delta grows. The present position of the delta allows deposition directly into Ballenas Basin by both suspended and bottom transport. Unfortunately, it is not known for how long this has been proceeding. It may be that, because the delta is now building into the basin from a position along its side, the present rate of sedimentation in the northwestern area is greater than it was in the distant past when the H_1 layer was deposited. The age of the modern delta, from Mathews and Shepard (1962, p. 1432) is less than 11,000 years, but more than 7,300 years. Taking 10,000 years as an acceptable age, and the average total depth of

sediment in the deep basin area as approximately 200 meters (650 feet), the average rate of deposition is found to be two centimeters per year. This rate, calculated for Ballenas Basin, does not apply outside this basin since it is a site of rapid deposition. It may not apply to Malaspina Basin which, partly because of the constriction between it and Ballenas Basin, may have a separate depositional history. It is interesting to speculate that, if the same rate were to continue, in another 10,000 years time sediments would reach the 180 meter (100 fathom) contour, filling the basins and burying several more ridges. In a Norwegian fiord basin, Holtedahl (1965) has estimated sedimentation rates are 1.0 centimeter per year. However, no rivers of size comparable to the Fraser discharge into it.

Turbidites and Slumps in the Basin Sediments

The estuarine sediments over much of the study area, and particularly in the Ballenas Basin, have as their origin the Fraser River. That this is the main source of sediment supply for the area is demonstrated by the fact that sediments thicken toward the river mouth and thin away from it. Other streams no doubt bring material into the Strait but, while their role could have been more important in the past, the Fraser River is now the major contributor.

Sediment distribution from the river mouth must occur in at least two ways. A suspended sediment load provides a

rain of particles that travel far into the Strait and settle onto the sea bottom. If this was the only method by which sediments were distributed, the result should be a mantle of sediments covering ridges as well as the basins, the thickness being related to the depth of water column overhead. Low flat areas such as the col between the northwest and southeast parts of Sangster Ridge should be thickly covered, with a layer at least half as thick as occurs in the basins on either side of it. However, the basins have four to six times the thickness of sediment as the low neck of the ridge. Other ridge areas are bare of sediments, as are many of the sidewalls of the basins. On slopes of less than about 10 degrees there may or may not be a sediment cover. Steeper slopes are invariably bare but where sedimentary cover is absent there is no disturbance or unevenness at the base of the slope suggesting slumping or sliding has occurred to remove it. In fact, the basin floor commonly remains flat almost to the steep sides and appears to meet the side slope rather abruptly. In some places a trench or moat occurs along the base of the sidewalls, sometimes continuous for distances of several kilometers. Therefore, sliding of sediments off the basin sides does not appear to be an important process in filling the basins.

It is difficult to see how sedimentation from suspension can lead to flat basin floors or to sidewalls bare of

sediment. Of course, disposition of material in suspended transport is subject to vagaries of currents and currents are, no doubt, important in distributing fine material over the Strait. It is therefore possible that currents have kept the sidewalls bare, especially those at low angles that could support sediments but do not. The presence of trenches along the sidewalls may also be due to current action. Little is known of sub-surface currents in the basin area. They may be strong enough to erode trenches or at least to prevent deposition in them. But even with currents maintaining bare sides and distributing sediments over basin floors, it is doubtful that suspended sediments alone could fill the basins in the flat-floored manner recorded on the profiles of Ballenas Basin. The presence of many strong reflectors that start at the delta and gradually weaken with distance from it, extending for many kilometers with a low slope and a plane surface (Plate I), and which, in at least some places, appear to be contained within banked sides, all tend to support the existence of turbidite layers in the basin. Turbidity currents from the Fraser Delta which dropped their heaviest material first, near the base of the steep delta front, could lead to reflectors having strong reflection co-efficients where coarse material is present, but weakening as the coarser fraction thins out with distance from the source. Smaller turbidity currents would probably vanish, like some of the reflectors,

before the end of the basin is reached. Stronger currents, perhaps from higher on the delta, or charged with a greater load, would persist longer and reach the end of the basin. Turbidity currents tend to flow within low levees and leave a flat-surfaced area in their wake. The profiles show areas where internal reflectors terminate at low inclines near the basin edges. Good examples occur near position C on Plate X and position D on Plate XI. Turbidity currents also leave moats or trenches around obstructions (Hamilton, 1967) which could explain these features in Ballenas Basin.

In an attempt to determine if turbidites were present in the upper few meters of the basin sediments, two cores were obtained from the flat floor of Ballenas Basin at positions marked in Figure 5. No layering was visible in either core and sample analysis did not reveal any systematic trends with depth. However, only three intervals in each of the seven foot cores were analyzed. It is possible, too, that organic activity on the basin floors may have destroyed any gradation caused by turbidites. Sand content in the analyzed samples was one to seven percent. The presence of that much sand can be explained by bottom transport in the nature of density or turbidity currents, but some medium and fine sand is also present in the suspended load of the Fraser River (Johnston, 1921). Thus the cores obtained neither confirm nor reject the turbidity current concept in this basin.

In Malaspina Basin the upper 30 to 60 meters (100 to 200 foot) of sediment is more transparent than stratified, indicating the presence of homogeneous rather than graded sediments. Thus, recent turbidites are not suspected in this layer. Plates II to IV, which cross Malaspina Basin, show deeper sub-bottom reflectors similar to those of Ballenas Basin. Turbidity currents therefore may have occurred in the past. However, as the reflectors slope to the south, bottom transport is likely to have been in that direction as well. Since the north end of Malaspina Basin is unexplored, sources of turbidites in that direction are unknown.

Compaction of Unconsolidated Sediments

Compaction of the sedimentary layers has undoubtedly occurred and should be observable on the seismic profiles. Compaction can be recognized by sagging of bedding where it is not supported by underlying rock (Cone, et al, 1963). Therefore, the thicker the sediment, the greater should be the degree of compaction. Thus, if one assumed that a horizon was originally plane to the point where it abutts the rock sides of a basin, by measuring the difference in depth presently exhibited by the horizon where it is supported by rock edges and in the central areas where it is not, an estimate of compaction can be made. Several difficulties are obvious in attempting such a measurement. The upper layers of sediment tend to smooth out the depressions left

by previous compaction. The top layers are therefore poor indicators of total compaction. As well, in the delta area the present upper layers are fore-set beds and are sloping rather than flat-lying. In Ballenas Basin, the top of sediments sometimes rests far up the sides of the rock walls or else turns down into a moat at the edges. These are not areas where compaction can be measured with reliability. Hyperbolic reflections from steep rock sides tend to obscure rock-sediment contacts of many otherwise suitable sites so that the top of sediments at the sides cannot be seen.

Despite the above difficulties an estimate of the degree of compaction was made in Ballenas Basin using the profile of Plate I. The northwest end of the lower H_1 layer is seen to be draped up the end wall of the basin and over various bedrock humps along the bottom. Subject to the assumption previously made, that the horizon was originally plane and horizontal, the H_1 sediments there have settled some 45 meters or about 30 percent of their original depth of 135 meters in this area.

III. ELEVATED AREA OF RIDGES

Along the mainland side of the Strait an area elevated above the basin floors extends from Thormanby Islands in the northwest to Burrard Peninsula in the east. Forming an upland upon which are situated two higher ridges, the

area increases in width from $2\frac{1}{2}$ kilometers near Thormanby Islands to a maximum of 13 kilometers before decreasing again toward Burrard Peninsula. West of Point Grey, Queen Charlotte Trench, 240 meters (130 fathoms) in depth, cuts through the elevated area into Howe Sound. However, Burrard Peninsula and Burrard Inlet to the east are so well related morphologically and geologically to the ridge and bank areas to the west that they are included in that area, although not specifically a part of the present study area.

The two afore-mentioned ridges extend in a northwest-southeast direction along the elevated area. McCall Ridge, nearest the mainland coast, is longest and covers much of the elevated area. The southeastern toe of this ridge may extend under the Fraser Delta sediments in a direction parallel or sub-parallel to Burrard Peninsula. Halibut Ridge, on the edge of the elevated area, is also long and narrow, although smaller in area than McCall Ridge. Both ridges support banks that rise to within a few meters of the sea surface.

A relatively flat basin area, called here Sechelt Basin, extends along the base of the mainland slope at a depth of 170 meters (92 fathoms). The flatness of the sea floor marks this basin as an area of thick sediments. In the west the basin drops to a terrace level below 200 meters (110 fathoms) at the end of McCall Ridge. In the east, a deep canyon is incised from the basin level to the floor of Queen

Charlotte Trench 75 meters (40 fathoms) below. A remarkably deep hole occurs in the floor of this canyon. Sounded to a depth of 292 meters (160 fathoms) the hole, called Jones Deep after a fabled character of the sea, has extraordinarily steep sides for an area of rapid deposition. Mean water depths surrounding the hole are 200 meters (110 fathoms). The fact that it is there at all requires explanation. The hole is discussed further in a later section.

Three other ridges exist in the northwestern Strait. Their different topographical outlines suggest at least some basic differences in their structure. Round Ridge, the most westerly on the slope north of Malaspina Basin, is, as its name implies, a round knob which rises to approximately the same height as the elevated terrace near it. Farther east on the same slope, South Ridge, actually two separate peaks of limited height, projects from the base of the slope near the floor of Ballenas Basin.

Sangster Ridge, a broad, nearly flat topped ridge separating Malaspina and Ballenas basins, extends in an almost east-west direction rather than in the usual southeast-northwest direction of the other ridges. A low neck, or col, connects the ridge to an eastern section consisting of several small peaks rising from a low base.

Unconsolidated Sediments

Plates II to XIV cross the elevated area and should be referred to in this section. A mantle of unconsolidated sediments covers much of the area. Near the Fraser Delta the layer is thick. In fact the southernmost part of McCall Ridge (Plate XII) is completely buried by deltaic sediments. Queen Charlotte Trench is also covered with more than 180 meters (600 feet) of similar sediments. Away from the delta, the sedimentary mantle thins except in the low areas between ridges and in Sechelt Basin. West of the highest points of McCall and Halibut ridges, ridge sediments are thin or absent over many areas, even where the ridges are flat. In Sechelt Basin unconsolidated sediments range in thickness to 60 meters (200 feet) or more. The basin sediments are flat-lying as far northwest as Mission Point, after which they thin considerably and show much relief as the basin drops to a lower terrace level.

Unconsolidated sediments on the ridges and in Sechelt Basin are mainly of the 'transparent' type, marked H_t on the line drawings. Since they become progressively thinner away from the Fraser Delta area, this is their obvious source. The ridge sediments join laterally with deltaic sediments of the H_2 and H_3 layers close to the delta (Plates XI and XIII). Where this occurs, weak stratification is visible in the H_t layer, possibly indicative of heavier grain-size material

closer to the source. The H_t sediments are seismically homogeneous, probably hemipelagic sediments brought by the Fraser River to the Strait where they have settled from suspension. Some material may come from other rivers and streams entering the Strait, or from inlets such as Howe Sound which have large rivers at their head but, compared to the Fraser River, these sources are negligible.

The manner in which transparent sediments react to sound waves is worthy of mention. A recent continuous seismic profile by R. Stacey (Dominion Observatory, personal communication, 1968) in which only frequencies less than 80 Hertz were recorded, barely detected the transparent layer, even though it is thirty or more meters thick. The profile of Plate X over Sechelt Basin, recording up to 300 Hertz, shows the transparent layer as a strong signal. However, its multiple, which should occur at twice the travel time or at approximately 0.440 seconds between positions E and F, is not present at all. A multiple at 0.5 seconds corresponds to the first buried sediment-sediment interface or unconformity reflected from the sea floor. This multiple carries the irregularities of the unconformity as proof of its origin. A multiple of the same interface reflected from itself rather than the sea bottom lies at 0.55 seconds. There is no multiple of the sea floor. From these facts it may be deduced that the transparent layer must have a greater

reflection coefficient for high frequencies, while low frequencies apparently are transmitted through the layer rather than reflecting from it. Coring indicates the transparent sediments are very soft, with a high percentage of water and possibly a 'soupy' sediment-water interface rather than a firm surface. Samples of the blue-grey mud ooze through the fingers with little pressure. Corers dropped into the layer return to the surface with the appearance of having been buried at least as deeply as the shackle on top of the core barrel. After drying, samples of these sediments shrink to about one half their original size.

Beneath the transparent H_t layer a strong reflector marks a second unconsolidated layer over the elevated area, marked H on the line drawings. It is present mainly under Sechelt Basin and in low depressions such as the valleys between ridges. The layer is thin and relatively flat surfaced and, like the lower H_1 layer in Ballenas Basin, has smoothed out much of the irregularities of the bedrock surface upon which it rests. Its occurrence, being confined to depressions or low areas, is somewhat patchy. A similar lower layer buried under more recent sediments is evident in many sedimentary pockets on the island slope as well. The general presence of an older layer in the Strait indicates some change in sedimentation of the Strait. No diastem is known, but a change in pattern of sedimentation occurred

in recent times when Fraser River sediments overflowed Fraser Ridge directly into Ballenas Basin. Before that time, sediment transport was apparently via a route south of the ridge. By entering at the side of the basin, the path to any point in the north and western Strait was decreased, thus the sedimentation rate, as well as particle size settling in the area, may have increased. Sediments of the lower unconsolidated layer are almost always flat-lying and found only in basins or depressions. Perhaps currents at the time of their deposition were strong enough to remove material from areas other than depressions.

Echograms and CSP profiles over South Ridge show that, although near the Fraser Delta, the thick sedimentary cover of nearby areas is also mainly lacking. No sediments are observed anywhere on Round Ridge while on Sangster Ridge, sediments occur mainly on the low, flat-topped col between the main ridge and its eastern toe. Although the col is just 30 to 50 meters above the basins on each side, sediments reach only 15 to 30 meters (50 to 100 feet) in thickness. Most other areas of the ridge are bare.

Pleistocene Sediments

A strong unconformable horizon marks the base of unconsolidated sediments. Beneath this horizon several seismic events, in what are believed to be Pleistocene sediments, are evident on the profiles. The most striking of

these underlies Halibut and McCall Ridges and the intervening low area. Several hundred meters of well stratified reflectors, identified on the line drawings by the symbol P_{mr} and given the name McCall Ridge unit, are mainly horizontal or dipping slightly to the west with gentle folds. They compose a large part of the sub-bottom ridge structure. Stratified areas are interspersed by areas of differing seismic character marked P_n. These show only limited stratification or none whatsoever. Some appear completely chaotic in structure. Correlation of the P_n areas is not usually possible from record to record because of rapid changes or fadeouts in both vertical and lateral directions, but the well stratified McCall Ridge unit provides a good marker that can be recognized over many profiles. While individual strata cannot be correlated, the top and bottom of the unit can be picked with a reasonable degree of certainty. The P_{mr} unit mainly overlies deeper bedrock whereas areas of non-stratified or chaotic P_n reflections occur upon or between the stratified unit. This sequence is well illustrated in Plate VII. Here, stratified reflectors underlie the terrace area north of Malaspina Basin. In an overlying P_n unit, events are either chaotic or unevenly and weakly stratified. A P_n unit rests under the peak of McCall Ridge on this profile. Most of the elevated area and the ridges upon this area are underlain by similar sections. The well stratified McCall Ridge unit

can be correlated from profile to profile from Queen Charlotte Trench to at least as far northwest as the end of McCall Ridge, a distance of 42 kilometers. From there to Thormanby Islands, a distance of five kilometers, bedrock possibly outcrops in many places with the P_{mr} unit between. The few profiles over the outcrop area are insufficient to show exact detail, but elongate topographic highs along the terrace edge (Plates III and IV) are, like Round Ridge and South Thormanby Island, probably of bedrock origin. The P_{mr} unit continues to underlie the terrace west of Thormanby Islands as well as a reef which continues to the north of these islands (Plate II). Based on the sea floor contours in that direction, it may extend as far as Bjerre Shoal, seven kilometers beyond the study area.

The age of the ridge structure below the unconsolidated sediments but above the bedrock is almost certainly Pleistocene. Pleistocene ridges trending in the same direction and having the same general configuration occur on land. Burrard Peninsula, for instance, is a Pleistocene ridge overlying Tertiary bedrock. It ends in sea cliffs not far from McCall Ridge. Pleistocene tills, drift, outwash and other glacial and inter-glacial features are common throughout the Strait of Georgia region, including some of the shorelines around the study area. Some of these sediments are unstructured, others, such as at Point Grey, are

well stratified and flat-lying. Profiles that pass or terminate near known Pleistocene shorelines usually record seismic structure with these characteristics. The north end of Thormanby Island nearest the reef is underlain by a thick sub-aerial section of stratified Pleistocene sediments. These sediments are well exposed on a 60 meter (200 foot) cliff overlooking the reef. Similar sediments occur on Savary Island and other large islands in the northern Strait of Georgia (Bancroft, 1913). They are also found on Vancouver Island (Fyles, 1963) and on the mainland (Johnston, 1923; Burwash, 1918). As mentioned previously, they outcrop on cliffs at Point Grey, Burrard Peninsula (Armstrong, 1956) not far from the eastern end of McCall Ridge. Their common occurrence and widespread distribution has led to speculation that at one time they completely covered much of the Strait of Georgia (Bretz, 1913; McConnell, 1914). It is therefore not surprising to find sediments under the submerged ridges whose morphological and geophysical character is suggestive of those found under sub-aerial conditions. Dredging and grab sampling in areas of outcrop on McCall and Halibut Banks have recovered angular to sub-rounded pebbles and cobbles, many with glacially carved striations. A Pleistocene age for the ridges is therefore a reasonable assumption.

As suggested above, it is possible that the McCall Ridge unit is composed at least in part, of sediments similar

to those on Thormanby and other islands. It probably predates at least the latest, or Vashon, stage of Cordilleran glaciation. The thick overlying chaotic events may then be till or unstratified drift from later glaciation. A sequence of till-stratified interglacial sediments-till occurs over nearby land areas, (Clapp, 1914; Fyles, 1963). The ridges may also be much older Pleistocene structures. In fact, since Vashon sediments on land are rarely greater than 30 meters (100 feet) thick and commonly less than 13 meters (40 feet) thick, the overlaying material may consist of much older till or drift.

The thickness of the McCall Ridge unit varies considerably over the area of ridges. North of Thormanby Islands, it approaches 500 meters (1,650 feet) in thickness. If the Pleistocene cliffs above sea level are included as part of this unit, 560 meters (1,850 feet) of section is represented. Toward the southeast, the thickness is less, but still considerable. Except for the area south of Thormanby Islands where the unit seems to disappear almost entirely, the thickness is nowhere less than 150 meters (500 feet).

Two other occurrences of similar seismic reflectors are tentatively identified as part of the McCall Ridge unit. The first, on the east side of Lasqueti Island slope, occurs as a thick, though possibly small layer directly across from the same unit off Thormanby Islands. In fact, the impression

gained from the one profile over the area, Plate II, is that the reflectors were at one time continuous across Malaspina Basin. Individual reflectors, however, cannot be matched across the basin.

A second small area on the south side of Sangster Ridge (Plate IV) also represents a thickness of several tens of meters. It is not present on adjacent profiles from either side and therefore must be limited in lateral extent.

At one time the P_{mr} layer may have been much more extensive than at present. The truncated strata on ridge sides, its great thickness and occurrence on both sides of Malaspina Basin point to this possibility. Although presently found mainly on the mainland side of the study area, the occurrence on the south side of Sangster Ridge suggests that it may also have existed in Ballenas Basin. Later erosion, perhaps by glaciers, may account for its removal.

Other ridges are not underlain by the same well-stratified P_{mr} reflectors. Some stratification does appear at South Ridge but it is deformed and not continuous. This ridge may be older than the McCall Ridge unit but relationships are not certain because of many unconformities in the Pleistocene record between the two ridges.

Round Ridge, like other bedrock areas, has no internal seismic character at all. Even a gas-exploder profile across it shows no internal structure. No Pleistocene

sediments appear on it. Its connection with surrounding bedrock areas is not known but it is suspected to be a granitic plug or boss similar to those found on Texada Island (McConnell, 1914).

The eastern end of Sangster Ridge and the col connecting it to the main body of the ridge are, below the unconsolidated sediment cover, seismically similar to Round Ridge. No coherent internal seismic character is apparent. However, under the main western part of the ridge, a continuous horizon does appear deep in the structure, dividing it into upper and lower layers. Few reflections occur above or below this horizon and it is not known if the horizon marks the top of bedrock or is an event in bedrock. The P_{mr} unit occurring on Sangster Ridge (Plate IV) is overlain by, and therefore older than, the layer above the horizon and on this basis the upper layer, which includes much of the ridge, has been assigned a Pleistocene age.

Dredging on Sangster Ridge has provided no freshly fragmented bedrock particles despite the fact that the dredge wire pulled taut several times as the dredge caught on the bottom. Mud, very fine sand, pebbles, some coated with manganese dioxide, and angular cobbles were dredged. The bottom appears to be very irregular judging from the way the dredge was continually snagged as it dragged along the sea floor. The ridge may be, in part, morainal in origin.

Lateral moraines of glaciers moving southeast on each side of Texada and Lasqueti Islands may have joined as a medial moraine, depositing large quantities of debris on the ridge site.

Bedrock

Underlying the unconsolidated sediments of Sechelt Basin and the Pleistocene deposits of the ridges, bedrock reflectors dip off the mainland slope toward the central basin. The reflectors are truncated by the erosion surface of the bedrock. Stratification terminates just offshore all along the mainland coastline. The bedrock of the mainland slope itself is unstratified. The stratified bedrock is designated by the symbol B_t and occurs from Queen Charlotte Trench to the end of McCall Ridge southeast of Thormanby Islands. Deep under the Pleistocene units of the elevated ridge area, the stratification cannot be seen, but the horizon marking the top of bedrock can still be followed toward the level of bedrock beneath the basins. Whether or not the horizon continues to represent the same event is unknown. It is therefore marked with the letter 'B'. Although a definite contact cannot be seen on the profiles, the stratified bedrock almost certainly overlies the crystalline rocks of the mainland slope. These latter rocks, mainly Coast Range intrusives extending from Howe Sound to the northwest, are either not penetrated by the seismic signal or show no

internal reflectors. By projecting the dips of the overlying B_t strata, an apparently unconformable contact results under Sechelt Basin. In the southeast a previously mentioned canyon occurs at the contact with the mainland rock on one side and truncated strata on the other (Plate X and Figure 8). The stratified bedrock, overlying as it does probable Coast Range intrusives, is possibly related to the Burrard-Kitsilano Formations of Late Cretaceous-Early Tertiary age near Vancouver. Sub-aerial bedrock exposures of these formations outcrop near Kitsilano and other areas on Burrard Peninsula. They dip to the south at eight to ten degrees and are overlain unconformably by thickening deposits of Pleistocene age. On parts of the southern slope, these exceed 215 meters (700 feet) in thickness (Armstrong, 1956). Similar conditions are evident under McCall Ridge (Plates VIII, IX and X), with strata dipping off the Coast Range intrusives and covered by Pleistocene sediments thickening to the southwest. The difference in structure between the undersea ridge and the sub-aerial peninsula is mainly in the layer of unconsolidated marine sediments covering much of the former.

The stratified bedrock along the mainland side is seismically similar to bedrock strata (designated B_k) along the Gulf Island slope. Bedrock of both areas shows similar internal stratification. How much of the bedrock floor of the Strait of Georgia is underlain by these reflectors and

to where they go on the Vancouver Island side west of Gabriola Reef is not known from the data available. A diligent search of the profiles gives some indication of the presence of internal bedrock stratification under various parts of Malaspina and Ballenas Basin, but lack of continuity prevents positive correlation. The bedrock under the basins may well be Tertiary as it is under Whatcom Basin in the Fraser Lowland. At Fraser Ridge, for example, weak stratification under the deepest horizon may represent the top of Tertiary (?) bedrock (Plate XXIII). The buried eastern section of Finger Ridge (Plate XVII) shows an anticline evidently formed in the same strata. The gas exploder profile of Plate XIX also indicates an anticline in bedrock west of Fraser Ridge. However, the record is difficult to interpret.

South Ridge could also be stratified bedrock but the upper section of at least one, and perhaps both, of the two peaks is likely Pleistocene sediments.

The lower layer of Sangster Ridge is possibly bedrock, but of what age or relationship to other bedrock of the area is not known. No correlatable stratification or other identification is apparent. The bedrock exposures on the edge of the elevated terrace area north of McCall Ridge may be older rock related to the pre-Upper Cretaceous on South Thormanby Island. Small topographical features on the

nearby elevated terrace area and on the mainland side of Malaspina Basin are seismically unstructured bedrock. Round Ridge is similar in character and no doubt related. Since the eastern toe of Sangster Ridge is bedrock of the same seismically unpenetrable character (Plate VII), these features may be a continuation of the pre-Upper Cretaceous arch that separates the Nanaimo Lowlands into two basins at Nanoose Bay. More detailed work in this area may show if that is the case.

IV. ROBERTS SWELL AND THE NEARBY MAINLAND SHELF

South of the 49th parallel a smooth dome-shaped topographical high lies at the foot of the mainland slope off Point Roberts. Extending almost fifteen kilometers across the Strait toward the Gulf Islands, this area, called Roberts Swell is, geomorphically and on the basis of other evidence to be presented, not a part of the present Fraser Delta. Its surface in most places is smooth and featureless with a gentle slope to the southwest and west. On the southwest side it is bounded by the long linear U-shaped Trincomali Trough, and on the southeast by Boundary Basin.

The mainland slope above Roberts Swell differs north and south of Point Roberts. To the north the slope angle is slightly less than the upper delta front. However, at Point Roberts the slope steepens considerably. Water depths drop

almost immediately offshore to 110 meters (60 fathoms) with no evidence of a shelf. For several kilometers southeast of Point Roberts the slope is formed by Roberts Reef, whose western side is steep and irregular. South of the end of the reef, the mainland slope becomes smooth and regular, angling more gently downward to the floor of Boundary Basin. This slope characteristic continues as far as Alden Ridge, a structure partly buried by the mainland slope sediments. Like Roberts Reef, Alden Ridge forms a structural barrier to shelf sediments, ponding them on its eastern side, but there is no evidence to suggest that the ridge is a part of, or similar to Roberts Reef. A wide shallow bay, Boundary Bay, occurs to the east of Roberts Peninsula and Roberts Reef, on the mainland shelf.

Plates XXIV to XXX show profiles over Roberts Swell, including one gas exploder record obtained by Shell Canada Limited. Five profiles run transversely across the Strait; one profile, Plate XXX, is perpendicular to the others, running along the axis of the Strait.

The Roberts Swell Unit

The profiles show the smooth and relatively flat-topped Roberts Swell with prominently steepened sides on the southwest, south, and southeast. In the northwest, the surface of the swell slopes down at a low angle toward Ballenas Basin and merges with the lower delta slopes.

Penetration achieved is much better than on the upper delta. Instead of the internal roughness and random stratification common on southern delta profiles in equivalent water depths, the records show a weak but smooth internal stratification under Roberts Swell. The reflections mark the Roberts Swell unit, denoted P_{rs} on the line drawings. Stratification is mainly conformable, at least under the smooth top, with the sediment-water interface. Little internal stratification is present at depths greater than thirty to sixty meters below this interface. That energy does penetrate through the sediments is shown by the underlying units, marked P_s, recorded below the base of the P_{rs} sediments. Lack of strong reflections in the P_{rs} unit indicates the sediments are almost seismically homogeneous throughout the section.

Roberts Swell is separated from the Gulf Island slope by Trincomali Trough. Sub-surface reflectors underlying the top of the swell in this region remain conformable to the sediment surface, even as it curves into the central region of the trough. Stratification turns down and terminates abruptly against the rising bedrock of the island slope (Plates XXVI, XXVII and XXVIII). Roberts Swell sediments occur under only part of the trough leaving at least the western half floored by bedrock. The Roberts Swell unit is not eroded here.

Farther south, sub-bottom stratification of Roberts Swell is truncated by the side of Trincomali Trough as it curves to the east into Boundary Basin (Plate XXIX). In this area the trough, which has cut deeply into the sea floor, has established itself in Roberts Swell sediments at the bedrock contact. Although the trough flattens out to the east and disappears in Boundary Basin, truncation of bedding remains characteristic of the southern and southeastern borders of Roberts Swell (Plate XXX). Here, a scarp formed by the ends of the truncated bedding stands at angles of five degrees or more for heights of 30 to 60 meters (100 to 200 feet). Internal reflectors of the Swell, which here dip gently to the northwest, are truncated at the scarp or on the top of the swell above the scarp. The scarp marks the limit of Roberts Swell as a morphological feature, but deeper stratification in the Roberts Swell unit can be traced into Boundary Basin to the south. The unit therefore extends beyond the limits of the swell in that direction. The sharp uprise of a bedrock feature under the scarp (Plate XXX) is due to the ship's track angling up the side of a ridge buried under the eastern part of the swell. The ridge is shown clearly in Plate XXXI.

Southeast of Roberts Swell and contiguous with it, a thickness of approximately 140 meters (450 feet) of stratified sediments lies at the base of Roberts Reef and extends

at least part way under Boundary Basin before thinning out and disappearing (see Plate XXXI). Stratification is mainly horizontal with gentle local dips. These sediments, too, show evidence of some truncation by the present sea floor. This fact, coupled with the smoothness of the sea bottom contours joining this area to Roberts Swell, and the known extension of Roberts Swell type sediments into Boundary Basin, relate these sediments to the P_{rs} unit.

Despite the proximity of the river mouth and the presence of overlapping delta sediments, the seismic profiles show no evidence of any delta sediments (H_3) overlying the flat, smooth top of the swell. Plates XXIV and XXV, over the northern part of this region, and Plate XXX, the longitudinal profile, do, however, clearly show upper delta sediments transgressing over typical Roberts Swell reflectors at the side nearest the delta. Where the overlap begins, there is a change not only in the character of the seismic signal, but in the sediment-water interface as well. The sediment surface changes from rough delta slope topography to the smooth top of the swell. There is also a marked decrease in the angle of slope at the base of the encroaching delta sediments. The smooth top, with a slope of only about four parts per thousand or fifteen minutes of arc, is completely uncharacteristic of the present delta front. Furthermore, internal reflectors marking stratification in the swell tend

to dip down and pass under the delta sediments. It is therefore clear that Roberts Swell sediments are not only different seismically from modern deltaic sediments but, because they underlie the delta, they must represent a pre-modern-delta episode of sedimentation.

Recent dredge hauls made up-slope between water depths of 210 to 140 meters (116 to 78 fathoms) on the southern marginal scarp of Roberts Swell show these sediments to be dark grey, muddly sands. Many irregular nodules and pelecypod casts of hard brown calcitic or phosphoritic material were recovered. Much broken shell material was dredged along with sub-angular cobbles and pebbles of igneous origin. Several chunks of poorly consolidated clinkers of sandstone were also found to be embedded with shells and shell fragments. If Roberts Swell type sediments are semi-consolidated, that would explain the better seismic penetration achieved there than on the delta where the sands are younger and less consolidated. Past sampling on the swell has provided the results listed in Table II, with an average composition of glacio-marine drift of the Fraser Valley for comparison. The sediments have a high sand and clay content. Mathews and Shepard (1962, p. 1492) postulated that the high incidence of sand south of the Fraser River may be due to either relict sands, or to a gating action of bottom currents in the river by tides. The incoming flood tide, which sends a

saline wedge of marine water under the fresh river water, was thought to check the bottom current and thus the bottom transport of coarse grained material so that it does not reach the river mouth during the flood stage. During ebb tide, the coarse grained material was suspected to flow freely from the river to the south. But with this explanation sediment deposition should be heavy south of the river. The seismic records indicate the opposite is true. No Fraser River sediments are settling upon the top of Roberts Swell.

TABLE II

SEDIMENT GRAIN-SIZE ANALYSIS
IOUBC Data Report No. 20, 1962

Station No.	Latitude	Longitude	Depth (Fathoms)	Sand %	Silt %	Clay %	Median Diameter (phi units)
3	48°51.5	123°07.0	87	42	39	19	4.3
5	48 55.7	123 13.1	77	52	34	14	3.95
167	48 55.5	123 06.8	65	58	26	16	3.5
171	48 59.7	123 12.2	63	59	18	23	3.45
Glaciomarine drift of the Lower Fraser Valley (After Armstrong, 1957, p. 4)				40	50	10	

It is known that tidal currents are strong over the swell. Table III, from Pickard (1956), lists current measurements taken with an Ekman current meter suspended sixteen inches above the bottom in a line from Roberts Peninsula to Galiano Island. Highest current velocities occur along the

mainland slope, but currents at other points over the swell are also high.

TABLE III

BOTTOM CURRENT MEASUREMENTS ACROSS ROBERTS SWELL
Gabriola Island to Roberts Peninsula

	Latitude	Longitude	Depth (Fathoms)	Bottom Currents (knots) Maximum Flood	Maximum Ebb	Mean Flood	Mean Ebb
D	48°55.4	123°23.5	23	0.3	0.35	0.1	0.15
E	48 55.7	123 21.8	84	0.7	0.6	0.25	0.15
F	48 56.3	123 19.9	105	0.8	0.4	0.35	0.15
G	48 57.6	123 14.5	78	0.4	0.6	0.15	0.2
H	48 59.3	123 08.4	54	0.95	0.7	0.5	0.3
J	48 59.6	123 07.0	23	0.3	0.35	0.15	0.15

(After Pickard, 1956)

Surface circulation in the Strait is, in general, counter-clockwise with a predominant northward movement along the mainland side (Waldichuk and Tabata, 1955). If the same circulation holds near the bottom of the Strait, main sediment transport should be to the north along the coast, with a consequent reduction in sedimentation in the Roberts Swell area. This is in agreement with the continuous seismic data.

The high percentage of coarse fraction found on Roberts Swell and to the north could be due to winnowing by tidal currents resulting in a lag deposit of coarser material being left behind while the lighter material is swept farther away. That erosion does, or did, occur at Roberts Swell demon-

strated by the presence of truncated bedding on the sea floor, pointed out previously at the southeastern margin of the swell. With the current velocities measured, erosion of the surface of the swell is possible and may still be occurring. Sand and coarse material from the swell could then be swept north along the mainland slope by a counter-clockwise current. This would account for heavy sand concentrations south of the main river mouth.

Thickness of the Roberts Swell unit. Only one profile, the gas exploder record of Plate XXVII, penetrates to the base of Roberts Swell sediments over a complete traverse. On this profile, greater stratification is observed through the depth of sediments because of the greater output power of this equipment. The configuration of the basin containing the sediments is well marked. Most surprising is the extreme thickness of the Roberts Swell unit. Assuming a velocity of 1,800 meters per second (6,000 feet per second), perhaps a conservative estimate, a maximum thickness of 455 meters (1,500 feet) is obtained. An average thickness of 390 meters (1,250 feet) is maintained over a distance of eight kilometers or more across the Strait. An area of more than 100 square kilometers is covered by the sediments. The total volume

may therefore be of the order of 40 cubic kilometers, not including those sediments outside Roberts Swell which are identified with the P_{rs} unit.

Age and source. The Roberts Swell unit overlies and therefore post-dates both Roberts Peninsula and Roberts Reef. Sea cliffs on Roberts Peninsula are composed in part of sediments of the Olympia Interglacial episode immediately preceeding the last major glaciation. The Roberts Swell unit is therefore younger than the Olympia Interglacial. A pre-modern delta age is imposed by the present seismic evidence and confirmed by the finding of angular cobbles with glacial striations in dredge hauls. Cobbles of the size and nature found could only have arrived in that location by ice transportation. Whether the cobbles are part of a glacio-marine till or a cobble pavement from drift ice is not immediately apparent. The absence of glacial features on the surface of the smooth swell and within its internal structure, as opposed to the irregular character of Pleistocene structures nearby, suggests that deposition occurred after the departure of the latest Cordilleran ice sheet from the Strait. That is, the sediments are most probably post-Vashon, but pre-modern-delta, in age. Floating sea ice must have been abundant during the later Sumas Stade of valley glaciation which may account for the cobbles over the swell. Carbon-14 dates on weathered-appearing shells dredged with the cobbles

gave ages of 1,845±160 years B.P. and 6,415±280 years B.P. The sediments, and the scarp itself, must be older than this.

The surface slope of sediments and the dip of internal stratification indicate the source direction was to the southeast, possibly in the Boundary Bay area. Glacio-marine and marine deposits of post-Vashon age are found in that area and throughout the Fraser Valley (Armstrong, 1956, 1957, 1960; Easterbrook, 1963). The Roberts Swell sediments are possibly related to the same post-Vashon events.

It is also possible to speculate that the Roberts Swell unit may be part of a former delta of the Fraser River. Armstrong (1960) suggests that the present course of that river was developed in the post-Sumas Stade. Before that time the river is believed to have passed through the Sumas Valley south of its present channel to enter the Strait in the Bellingham Bay area where the Nooksack River now discharges. If this is so, Roberts Swell sediments could represent the northern part of a previous delta. A seismic line paralleling the axis of the Strait but six or eight kilometers west of the present line would help to resolve this problem by providing information on the relationship of the swell to the lower and deeper sediments of the present delta, especially to the ancient bottom-set beds (H_1) that occur directly upon the bedrock floor.

Pleistocene Sediments

Roberts Peninsula is known to be composed of post-glacial, glacial, and non-glacial sediments of Pleistocene age. A deep well (Richfield Pure Point Roberts) near English Bluff penetrated through 258 meters (847 feet) of Pleistocene sediments before reaching Tertiary (Miocene?) at 199 meters (652 feet) below sea level. Therefore, near Roberts Peninsula a Pleistocene unit would be expected to overlie Tertiary bedrock in shallow water. At the closest approach of the survey ship to Roberts Peninsula, near Point Roberts (Plate XXVIII), the sea floor is an unconformable surface over which no deltaic or Roberts Swell type sediments occur. Reflectors are characteristics of those which in other areas are known or suspected to be Pleistocene drift deposits of the chaotic or non-stratified P_n type. Therefore it is reasonable to infer that, near Roberts Peninsula, the sea floor is composed of Pleistocene drift and till deposits similar to those found in cliffs along the shore. By correlation between profiles and by the reflection character on the records, the same sediments can be extended to include Roberts Reef. These are designated by the symbol P_n. Evidence of Pleistocene age would be strengthened if a Pleistocene-Tertiary contact, probably represented by an unconformity near 200 meters (650 feet) below sea level, could be found. Only three records show sufficient of the Roberts Peninsula area to be

able to look for such detail. From these profiles (Plates XXVII, XXVIII and XXIX) no definite conclusion can be drawn. A reflector does appear at about the proper depth, but it is not certain if it is actually the Pleistocene-Tertiary boundary or an internal event in the Pleistocene. The deep record provided by the gas exploder profile of Plate XXVII is too cluttered by multiples to show a definite horizon. The presence or position of Tertiary bedrock is therefore questionable in the seismic records of the Roberts Peninsula area.

Mainland Shelf Sediments

Information provided by the seismic profiles on the sub-bottom sediments of the Boundary Bay-Alden Bank mainland shelf area is interesting but not conclusive. Plates XXXI, XXXII and XXXIII illustrate the sub-bottom character of this mainland shelf area. Referring to these profiles, it is evident that the shelf is underlain by at least two sedimentary layers, and perhaps three. The geological history of the layers is not readily apparent. Thick unconsolidated sediments under the mainland shelf extend at least as far south as Alden Bank, the southern limit of the study area. Shelf sediments are contained east of Alden Bank and Roberts Reef structures and overlie their lower parts. Stratification is well defined and, in the main, conformable although

at least one shallow unconformity is present under the shelf on the most southern profile (Plate XXXIII).

Upper layer of shelf sediments. The upper layer of sediments, marked H_3 in the line drawings, covers most of the shelf area. In the Boundary Bay area, where it rests conformably over the layer below, it is thick, probably exceeding 110 meters (360 feet), but thins toward the outer shelf edge by truncation and thinning out of bedding (Plate XXXII). A few kilometers farther south, near Alden Ridge, (Plate XXXIII), the upper layer is no longer flat-lying, but dips toward the coast to rest unconformably upon the layer below. Although thinner than in Boundary Bay, sea bottom contours indicate it is probably the same sedimentary unit. On this profile its thickest section is only 36 meters (120 feet) and thins toward the coast as the layer below rises to the sea floor.

The sub-aerial Fraser Delta includes the north shore of Boundary Bay east of Roberts Peninsula. Johnston (1923, p. 40) reports a well just north of the bay penetrated about 122 meters (400 feet) of Recent delta deposits before reaching glacial till. Alluvial sands, silts, silty sands and silty clays were also found in more recent boreholes drilled in shallow water in Boundary Bay just north of the International Border. One borehole passed through 105 meters (344 feet) of delta sediments without meeting pre-delta material (Borehole

No. 2, Golder, Brawner and Associates). The horizon marking the base of the upper layer in Plate XXXII subtends the mainland slope at a depth of 130 meters (430 feet) below sea level, and can be traced to the northeast as a horizontal reflector for at least 5.5 kilometers (three miles) into the bay. If it can be extrapolated another 13 kilometers (eight miles) to the site of the boreholes it becomes clear that the upper layer under the shelf is a part of the modern delta of the Fraser River. Farther south on the shelf the upper layer may represent Recent Fraser delta sediments thinning out in that direction.

Middle layer of shelf sediments. Below the H_3 layer, a layer designated by the letter 'Q' underlies the shelf and slope. Flat-lying, conformable stratification in the layer is truncated by the mainland slope between Roberts Reef and Alden Ridge. A strong reflector marking its upper surface terminates at the edge of the mainland shelf (Plate XXXII) just at the top of the slope. East of Alden Ridge the layer thickens toward the coast and may outcrop on the sea floor near the eastern end of Plate XXXIII, about two kilometers from Point Whitehorn.

Several speculations are possible for the origin of this layer. Near Point Whitehorn coastal uplands are composed of Late Pleistocene till and glacio-marine drift (Easterbrook, 1963) which rest unconformably upon an irregular

and eroded surface of thick pre-Vashon marine sediments, the Cherry Point silts. This erosion surface appears at the base of sea cliffs south of Point Whitehorn and not more than five kilometers from the site of the profile of Plate XXXIII. This suggests that the middle layer may be correlated with the pre-Vashon Cherry Point silts under the erosion surface. However, since on shore these are overlain by till and a till layer cannot be identified on the seismic records (there are no chaotic reflectors or diffraction patterns usually characteristic of till), this is unlikely. Also, basic differences in the profiles of the Pleistocene sediments recorded near Roberts Peninsula which are both Vashon and pre-Vashon in age (Johnston, 1923) and those of the middle layer recorded near Point Whitehorn, tend to support an opinion that the middle layer here is not pre-Vashon. That the layer represents later Vashon glacio-marine deposits is also not well supported. The Vashon glacio-marine drift of the area is, according to Easterbrook (p. 1475) unstratified pebbly, sandy clay reaching thicknesses of only 20 meters (70 feet). The middle layer on the adjacent shelf is as much as 150 meters (500 feet) thick. If the lowest shelf layer, which is seismically similar and conformable to the middle layer, is included, the combined thickness probably exceeds 200 meters (650 feet). If the upland deposits did increase in thickness to seaward, reaching the amount required by the

seismic evidence, then the latter possibility may be more acceptable.

The well north of Boundary Bay reported by Johnston encountered clay and boulders, apparently drift, below 122 meters (400 feet). Below 146 meters (480 feet), blue clay was met, with successive deeper layers of clay and boulders. If the sediments overlying the middle layer are Recent delta material as found in the Boundary Bay well down to 122 meters (400 feet), then the middle layer in the Boundary Bay area could be the drift recorded by Johnston as lying under those sediments.

Since the upland areas along the nearby coast are separated by deep troughs now filled with several hundred feet of post-glacial alluvium (Easterbrook, 1963, p. 1475), another possibility may be that the shelf sediments are the seaward extension of this alluvium. Thus the middle layer as well as the upper one, especially in the Boundary Bay region, could be post-glacial sediments.

On the profile of Plate XXXII, the middle layer extends into Boundary Basin and overlies sub-bottom units of suspected Pleistocene age (P_n) in that area. Roberts Swell type sediments also overlie the same units. The Roberts Swell unit is found at depths below sea level equivalent to the middle and lower layers of the mainland shelf. Both Roberts Swell sediments and the middle layer are truncated

by the sea floor. These sediments are not directly correlatable on the seismic profiles but, from the above data, the inference is that they are related and may be parts of a common unit. If they are, the lower layers of the mainland shelf may then also be part of a former delta of the Fraser River when it discharged in this area. More detailed seismic profiling in this area would resolve this more fully.

The lowest layer of shelf sediments. Little seismic information is available on the lowest layer. The distinction between it and the layer above is based upon the presence of a strong conformable horizon between them and thus the distinction may be more apparent than real. The marker horizon could be an exceptionally strong internal reflector within a single depositional unit. The layer is therefore not designated separately from the one above it in the line drawing interpretations.

The layer lies over and therefore post-dates Alden Ridge. Its upper marker is flat-lying and the layer seems to appear only on the east side of the ridge. Reflectors at the same depths to the west of the ridge are mainly different in seismic character, especially the deeper ones. Thus the source of these sediments must have been to the east.

Due to the thickness of overlying sediments and noise and reverberation on the records the base of the layer is not clear. An indefinite reflector in Plate XXXIII at 0.34

seconds, equivalent to about three hundred meters (1,000 feet) below sea level perhaps marks the base. In nearby coastal regions bedrock below Pleistocene occurs at depths near this value.

V. BOUNDARY BASIN AND ALDEN RIDGE

The southern end of the Strait of Georgia is unique in the study area. Few linear trends are present in the structural geometry of the sea floor, nor are there any flat-floored areas as in the area of northwestern basins.

The triangular shaped Boundary Basin is bounded on its three sides by Roberts Swell, the mainland slope, and the island slope. The floor of the basin is broadly irregular in contrast to other basins whose floors have been smoothed by sediment deposition. Trincomali Trough enters the basin at the western apex of the triangle as a narrow U-shaped trough. As the basin floor widens, the trough curves to the east and north, following the edge of Roberts Swell and changing to a V-shaped canyon incised into the basin floor. Before reaching the central basin, it broadens out, flattens, and disappears. The deepest area of Boundary Basin occurs at Boundary Pass where a depth of 269 meters (147 fathoms) is recorded (Canadian Hydrographic Chart No. 3450).

Alden Ridge, on the mainland slope east of Boundary Basin, rises to within a few meters of the surface of the

Strait at Alden Bank. Water depths east of the bank are less than 75 meters (40 fathoms) whereas to the west they fall sharply to more than 180 meters (100 fathoms) in Boundary Basin.

The sub-bottom character is also unlike that of other basins in the study area. Seismic penetration is, in general, no more than one quarter second. Reflectors, including what appear to be major horizons, are not always continuous. Gaps in reflectors make correlations uncertain, as do rapid changes in sub-bottom character from one profile to the next. Even the profile along the axis of the Strait (Plate XXXIV) does not give much assistance in interpretation. Despite these problems, two, and possibly three, sub-bottom reflection characteristics of the basin indicate the compound nature of the region.

Sediments

An upper, horizontally stratified layer persists over much of the basin floor. Plate XXXIV, which is continuous with Plate XXX along the axis of the Strait, shows the layer continues under Roberts Swell, identifying the upper sediments with those of Roberts Swell unit and extending that layer to Alden Ridge in the south. Stratification is mainly horizontal but dips gently northwest, near Alden Ridge. A thicker section of the unit is banked against Alden Ridge forming a

flat-topped terrace. Sediments under the terrace are truncated by the sea floor. Stratification becomes less distinct in the mid-basin area.

Truncation by the sea floor of horizontal reflectors in the P_{rs} unit occurs on the mainland slope, at Roberts Reef, and at the southern margin of Roberts Swell, as well as at Alden Ridge. These occurrences, at very similar water depths, almost surround Boundary Basin and suggest that at one time the Roberts Swell type sediments were continuous over the basin. Thus, if the basin were refilled with these sediments to a depth of about 130 meters (70 fathoms), the result would join the truncated fragments into one integral unit.

Further support for this conclusion can be reasoned from evidence of the source direction of the P_{rs} sediments. On Roberts Swell the source direction was southeast, an impossibility if Boundary Basin existed at the time of deposition, but satisfactory if the basin was filled with sediments.

Considerable erosion must have occurred since deposition to create Boundary Basin. Up to 90 meters (300 feet) of sediments may have been removed from the sea floor. To change from a depositional state to an erosional state also attests to a large change in marine conditions after emplacement. Sub-aerial erosion requires emergence of 180 to 220

meters (600 to 800 feet), an amount not noticeably supported by other evidence in the area. Rather, as no deposition is presently occurring in the basin, it seems more logical to accredit the erosion to marine forces, possibly scour caused by large tidal currents entering the area from the channels to the south. Currents in Boundary Pass reach a velocity of 2.5 meters per second. A depression at the pass, most noticeable in Figure 5, may be indicative of the power of marine erosion in that area. Currents are also strong over Roberts Swell, as pointed out in Table III.

At a time during or shortly after the recession of Vashon glaciation from the Strait, the sea stood at a considerably higher level relative to the land. Various authors estimate relative sea level was 180 to 300 meters (600 to 1,000 feet) higher than at present (Easterbrook, 1963; Armstrong and Brown, 1954; Armstrong, 1956). The effects of tidal currents on the sea floor could therefore be vastly different with that much additional water overhead. Tidal velocities in the enlarged passes, as well as in the Strait, would be less. Sediment deposition may have occurred at such a time, with waning glaciation providing an abundance of detritus to build up large marine or glacio-marine deposits. As the land rose in isostatic adjustment to the melted ice load, currents along the sea floor no doubt increased in velocity and likely in turbulence as well, with increased

scouring effect. Post-glacial emergence may have exceeded the post-glacial transgression of the sea so that relative sea level was for a time, lower than at present (see Easterbrook, 1963, p. 1480). During a low stage of sea level, erosional effects could increase to the point where sea floor sediments in the southern Strait were vigorously removed. Later eustatic sea level rises overtaking isostatic emergence would then bring about the present state.

The base of the P_{rs} layer in central Boundary Basin is difficult to determine. Plate XXXIV, the axial profile, indicates the base lies near 0.5 seconds below sea level, a result entirely compatible with that of the gas exploder record of Plate XXVII over Roberts Swell. However, difficulties in interpretation arise just northwest of Alden Ridge. The axial profile of Plate XXXIV indicates mainly stratified material in the sub-bottom while transverse profiles indicate large areas of non-stratified sediments. The areas cannot be defined sharply on any single profile because the characteristics tend to grade into one another indistinctly rather than making a sharp contact. The areas are therefore identified only by their seismic character and marked P_n or P_s accordingly.

Since few continuous reflectors occur in these deposits, the general impression is that the deep sub-bottom is either almost homogeneous or completely random and chaotic

in nature. Sub-bottom of this type appears to be confined to the southern and southwestern part of Boundary Basin west of Alden Ridge. The material overlies the side of Alden Ridge in Plate XXXII, but may form part of the ridge top in Plate XXXIII. The two profiles are not clear in this respect. An unknown horizon, perhaps bedrock, appears under the layer. To the west near Boundary Pass (Plate XXXII), a thick stratified section, probably sedimentary, underlies the layer, but overlies bedrock of the island slope. It is not certain what this section represents either. During several periods of glaciation vast ice sheets occupied the Strait of Georgia and surrounding regions. Unstratified and semi-stratified till sheets were deposited over wide areas. Undoubtedly the southern Strait did not escape, and all these unknown deposits may be remnants of till, although they are different in seismic character from till in the northern part of the study area. West of Alden Ridge the thickness of this till (?) is greater than 200 meters (600 feet).

Bedrock Under Boundary Basin

The nature of the deepest horizon under Boundary Basin is difficult to interpret in view of the limited information. The depth of the horizon, generally at more than 0.5 seconds below sea level, puts it at the general depth of the bedrock floor of Ballenas Basin to the north. It is

therefore most probable that the horizon marks pre-Pleistocene bedrock in the area.

Reflections from the suspected bedrock horizon are of two characteristics. In the northern basin the horizon truncates strata, marked B, which have an apparent dip to the southeast (Plate XXXIV, position C). This strata is no doubt correlatable with similar strata folded into a synclinal ridge near Roberts Swell (Plate XXXI, position C). The top of the buried ridge is about 275 meters (900 feet) below sea level. Bedrock from the Gulf Island side (marked B_k) cannot be traced directly across to the stratified bedrock under Boundary Basin but, from Plate XXXI, it rests at the same depth and has similar seismic characteristics. The nearest deep well is Richfield Pure Point Roberts on Roberts Peninsula, about fifteen kilometers away. Tertiary is identified there at about 200 meters (650 feet) below sea level. The Tertiary rocks are more than 1,800 meters (6,000 feet) thick at the well site and overlie similar Cretaceous rocks. The deep stratified layer identified with bedrock may therefore be either Late Cretaceous rocks similar to those on the Gulf Islands or Tertiary formations as found under the Fraser Lowland.

In the southern Boundary Basin area the character of the bedrock horizon changes. The recorded horizon is marked by a rugged and irregular surface which is persistant through-

out the southern area. No penetration occurs through the rough surface. Where the change in bedrock reflecting characteristic occurs, there is also a change in depth to the horizon.

The presence of two different bedrock reflections at different depths may serve to mark a change in bedrock lithology in the southern Strait. If this is so, the stratified unit is most probably the younger. Plate XXXIV (between positions B and C) shows both types of horizon and indicates their obscure relationship, although in no profile is it any clearer. An unconformable contact between the two characteristic bedrock layers is suggested.

In the same plate, bedrock under Boundary Basin merges into a multiple of the sea bottom which prevents it from being traced under Alden Ridge. Thus the relationship of structures under the ridge to those in the nearby basin is not known. The transverse profile of Plate XXXII suggests that the core of the ridge is bedrock, whereas the profile of Plate XXXIII seems to require the ridge to be composed of semi-stratified and non-stratified P_n material. Plate XXXIV shows conclusively that at least some of the ridge is underlain by stratified materials tentatively identified as P_s type sediments, but a bedrock interpretation is not ruled out.

VI. THE ISLAND SLOPE

The island slope has two major trends in the study area. The junction of these occurs at Gabriola Reefs east of Gabriola Island. A change in slope morphology also occurs at this point.

Along the Gulf Islands south of Gabriola Reefs the island slope is broken by a system of southeast trending parallel ridges. Each successive ridge crest is lower toward the Strait. In the southern area the ridges are closely spaced. Some, near the Gulf Island shores, break the surface of the water as rocky reefs and islets. Tumbo Island is a sub-aerial extension of one of these. Locally, ridges are steep-sided with angles up to 23 degrees measured from the seismic profiles. Steeper angles may, and probably do, occur. The average steepness of the Gulf Island slope is, of course, less than 23 degrees. Farther north the ridges diverge from the coastline and become lower and less distinct. However, they continue to form the island slope as far north as Gabriola Reefs where the trend of the slope turns abruptly westward. There, the ridges stop.

North of Gabriola Reefs the Vancouver Island slope assumes a west-northwesterly trend. In contrast to the Gulf Island slopes, the slope here is relatively smooth, dropping quickly from the narrow shelf to the 365 meter (200 fathom) depths of the adjacent basin. Slopes as steep as 25 degrees

occur in lower, oversteepened portions. The average declivity is approximately ten degrees, steeper than the average slope in the south.

Just north of Gabriola Island a shallow ridge angles across the slope forming a valley, or wide canyon, behind it. Entrance Island is a sub-aerial part of the ridge. The valley is apparently suspended, or 'hanging', 180 meters (100 fathoms) above the sediments of the deep basin and 360 meters above the buried bedrock, perhaps indicative of the effect of glacial erosion in the Strait.

Most of the transverse profiles across the Strait show some of the island slope but only a few approach close enough to shore to show the upper part. This over-caution on the part of the ship's navigators could not be avoided but, judging from the quality of those few records which did reach the upper slope, the loss is not too significant.

Unconsolidated Sediments

Thick sediments occur on the island slope only in the area adjacent to the Fraser Delta. The lower slope across from the Fraser Delta is terrace-like, the bedrock surface being at such a low declivity that delta sediments (H_3) can rest thickly upon it. Plates XVIII to XXII show the terrace-like slope with many pockets of sediment trapped in basins between bedrock ridges which rise above the general terrace level. A flat horizontal reflector near the base of sediments

in most pockets probably correlates with the top of the H_1 layer under the modern delta facies. Sediments in some pockets reach 75 meters (245 feet) thick. Away from the delta, surprisingly little sediments occur in depressions and valleys on the slope. The sea floor is mainly composed of exposed bedrock.

Northwest of Gabriola Reefs, little sediment overlies the comparatively steep-sided bedrock slope. Sediments occur only on a few places which are sufficiently low-angled. A thin wedge of Ballenas Basin sediments extends up some of the less steep lower slopes. Others are apparently too steep to retain sediments. Thin sediments also cover the floor of the hanging valley near Gabriola Island.

Near Neck Point sediments in a second valley, which may also be a hanging valley, fill a small, semi-enclosed basin and may have overflowed onto the floor of Ballenas Basin ninety meters (fifty fathoms) below. Echograms of Ballenas Basin floor below the valley show irregular humps on the otherwise flat floor which can be interpreted as landslide features from the valley above. This is the only known occurrence of such features in the study area.

In one area and possibly two at the base of the island slope, possible Pleistocene sediments occur. Near the northwest end of Roberts Swell on Plates XXIV, XXV and XXVI, stratified reflectors (P_s) overlie bedrock but underlie

Roberts Swell sediments. Both top and bottom of the P_s unit are unconformable. The strata are nearly horizontal, and underlie most, if not all, of the western margin of Roberts Swell.

A second area is recorded only on the gas exploder profile of Plate XIX. Little internal stratification is evident and the layer is marked P_n (?). A small topographical expression on the sea floor (Figure 5 and Plate XIX, position B) probably defines the limited extent of the area involved.

Bedrock

The island slope has two differing topographical characteristics which allow a natural division of the slope into a northern and southern part. The change in topographical character occurs at Gabriola Reefs and is matched by a change in seismic character of the rock underlying the slope.

Along the Gulf Island slope the offshore ridges are composed of bedrock which, from their location, are almost certainly part of the thick Upper Cretaceous Nanaimo Group rocks. They have been marked B_k in the line drawings. Bedrock rises from under the delta sediments toward the outcrop areas of ridges higher on the slope. Stratification in the bedrock, characteristic of the Gulf Island slope, dips steeply to the east, but decreases in dip under the mid-Strait area. Strata is truncated on the slope and at the ridges, but

becomes more conformable to the bedrock surface under mid-Strait. A gentle anticline in bedrock occurs under mid-Strait in Plate XXI. Similar truncated strata dip to east and west off Finger Ridge (Plate XVII) suggesting that this ridge may also be, at least in part, a breached bedrock anticline.

Under the ridges on the slope many hyperbolic patterns occur, perhaps resulting from reflections off weathered or upthrust edges of truncated bedrock strata. These cover and obliterate or prevent identification of real strata which, on the nearby Gulf Islands, dip between 15 and 30 degrees to the east (J. Muller, unpublished map). Hyperbolae become asymptotic to 45 degrees when reflected from points under the ship's track. With a record exaggeration of 12, the difference in angles of 30 and 45 degrees as recorded on the profiles is only 3 degrees. With only short segments of strata recorded, these are not easily resolved. Thus, on the slope area, uncertainty remains if some of the observed reflections are hyperbolae or dipping strata.

The southernmost slope (Plates XXIV to XXXIII) narrows and becomes steeper. Because of hyperbolic reflection patterns it is impossible to pick out stratification in most places. Since the slope is steep, the stratification, if present, is likewise probably steep. South of Boundary Pass

(Plate XXXIII) no recognizable penetration into the bedrock of the island slope is achieved.

The general appearance of the offshore ridges and the truncated strata (where present on the slopes) give the impression that the slope was formed by a series of faults which dropped the eastern side to a lower level relative to the west. Muller (Geological Survey of Canada, unpublished map) records many faults in the coastal area of the Gulf Islands and has interpreted them in terms of block tilting.

The topographical character of the Vancouver Island slope north of Gabriola Reefs is different from the Gulf Island slope to the south. No ridges occur as off the Gulf Islands. Instead, the slope is, in general, steep and continuous. The lower part commonly shows oversteepening below the surface of the Ballenas Basin sediments. This is not present south of Gabriola Reefs.

Seismically the slope north of Gabriola Reefs is characterized by a lack of reflectors under the bedrock surface (Plates II to XII, XVI and XVII). Because of this apparent lack of seismic penetration, little information on the bedrock character can be deduced. Even a gas exploder profile (Plate VI), despite greater power, gives no internal reflections from under the bedrock slope.

The different seismic character of the island slope north of Gabriola Reefs to that in the south implies a differ-

ence in physical properties upon which seismic reflections depend. Regional bedrock geology (Figure 2) shows the Gulf Island area is completely underlain along the Strait by thick Nanaimo Group rocks of Upper Cretaceous age. The fringe of islands along the Strait, including Gabriola Island west of Gabriola Reefs, is composed of rocks of the Gabriola Formation, the youngest or top member, of the Nanaimo Group. Depths of Upper Cretaceous rocks are estimated by Clapp (1914) to average 2,050 meters (6,760 feet) in the Nanaimo area while on small offshore islands he estimated 3,000 meters (10,000 feet).

Northwest of Gabriola Island a high arch of pre-Upper Cretaceous rocks in the area of Nanoose Bay separates the Upper Cretaceous geosyncline into two basins, of which only the southern of the two, Nanaimo Basin, is in the study area. A change in seismic character could be expected when going from Upper Cretaceous such as at Gabriola Island to the older section. The Upper Cretaceous sandstone, conglomerates and shales can be expected to have different reflection qualities than the pre-Cretaceous rocks of altered basic volcanics, pyro-clastics and older meta-sediments. Thus a change could be expected in the profiles taken to the west of Gabriola Island compared to those along the Gulf Islands. The seismic records at Gabriola Island are, however, little different from those near Nanoose Bay or Neck Point where pre-Cretaceous

rocks outcrop on the coast. Little sub-bottom penetration was obtained in either area. Penetration into bedrock begins south of Gabriola Reefs where the change in topographic character and trend also occur.

To be sure that such was indeed the case, a short profile was obtained starting from near shore at mid-Gabriola Island and extending across the slope and into Ballenas Basin. The results, shown in Figure 6, confirm that the record generated off Gabriola Island is more characteristic of the Vancouver Island slope than the Gulf Island slope to the south.

Why seismic penetration should increase south of Gabriola Reefs instead of north of Gabriola Island is not known. Lithology of rocks underlying the slope is not expected to change there. Where the change is expected, the seismic character of the record is not affected. A possible cause of this behaviour may be the steepness of the topography, which, northwest of Gabriola Reefs, becomes generally steeper than the area to the south.

Of all the profiles over the island slope area, only two continue through inter-island passages into areas known to be surrounded by Upper Cretaceous rocks. Because of hyperbolic reflections and reverberation, these profiles (Plates XVIII and XIX, XXVI and XXVII) are not conclusive in determining if reflections characteristic of these rocks are

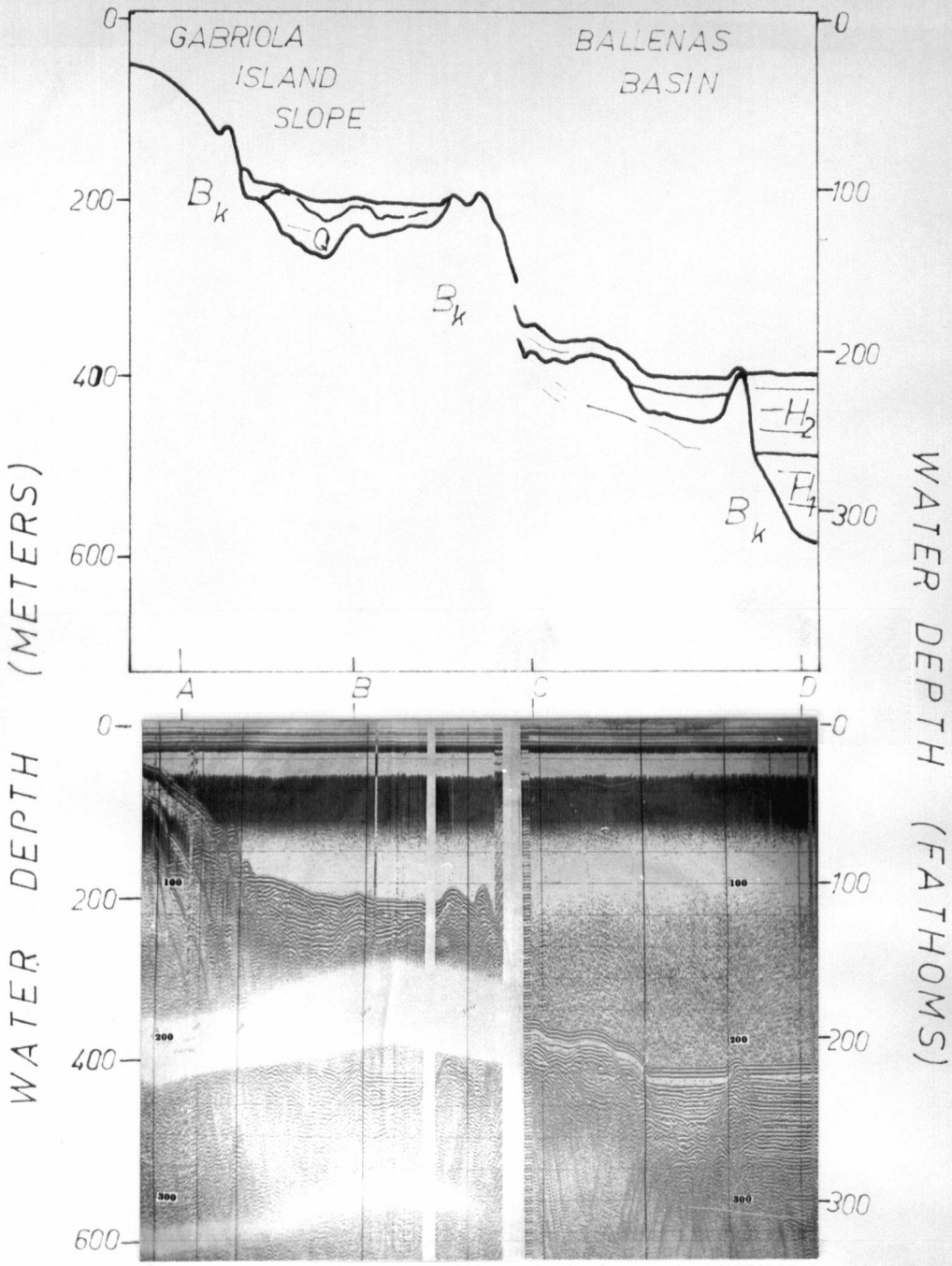


FIGURE 6. A continuous seismic profile extending northeast from nearshore Gabriola Island. Length is 8 kilometers. Positions on Figure 5.

continuous onto the lower slope. If they are, the lower slopes would almost certainly be part of the youngest of the Nanaimo Group formations. Strata on the island slope, if projected up dip toward the Gulf Island would, in the absence of faulting, overlie the islands by upwards of 300 meters (1,000 feet).

CHAPTER V

SPECIAL INVESTIGATIONS AND OBSERVATIONS

During the course of the study several small features became apparent on the seismic profiles that do not add greatly to the general view of the Strait of Georgia sub-structure but which are, nevertheless, interesting. The present chapter will deal with these.

I. JONES DEEP

One of the most puzzling features on the bathymetric chart of Figure 5 is a deep hole, called Jones Deep, on the floor of a submarine canyon leading from Sechelt Basin to Queen Charlotte Trench. Co-ordinates for the deep are Latitude 49°20'N, Longitude 123°28'.8 W. The canyon, cut into bedrock at the base of the mainland slope at the contact between the coastal rocks and the stratified offshore bedrock, is now filled with unconsolidated sediments of the 'transparent' type to a depth of several tens of meters. The elongate cone-shaped hole penetrates through the sediment to, or near, the bedrock base of the canyon which lies at a depth of 310 meters (170 fathoms) below sea level. This depth is 165 meters (90 fathoms) below the general level of surrounding sediments and 46 meters (25 fathoms) deeper than the sediment floor of nearby Queen Charlotte Trench. The

top of the hole is about one kilometer wide. Because of the spread of sound waves from the echo sounder the width of the hole at the bottom is not known, but at or near bedrock the hole is not likely to exceed a few meters in diameter. Side slopes average about 16 degrees but may steepen toward the bottom. Dredging in the hole obtained only soft blue-grey mud, typical of areas of transparent sediments. Echograms over the hole are shown in Figure 7. Seismic profiles of Plates XI, XIV and Figure 8 show the bedrock canyon with its covering of thick, unconsolidated fill.

Explanations for the existence of a deep hole in unconsolidated sediments must also account for its maintenance in an area of rapid sedimentation. Recent fault movement under the canyon, for instance, could result in a deep cleft but it does not provide a mechanism for maintaining the hole. Also, nearby profiles provide no evidence of recent motion in the surrounding rock or sediments. Fault motion would not be likely to leave soft muds at the high angles observed on the sides.

Currents in the water column may prevent sedimentation or even erode the sea bottom in certain areas, for instance, in Boundary Basin. Some ridge areas may be kept clear of sediment cover by currents, but it is doubtful that currents could excavate such a deep, steep-sided hole. Larger holes or trenches do occur in areas where persistent heavy, tur-

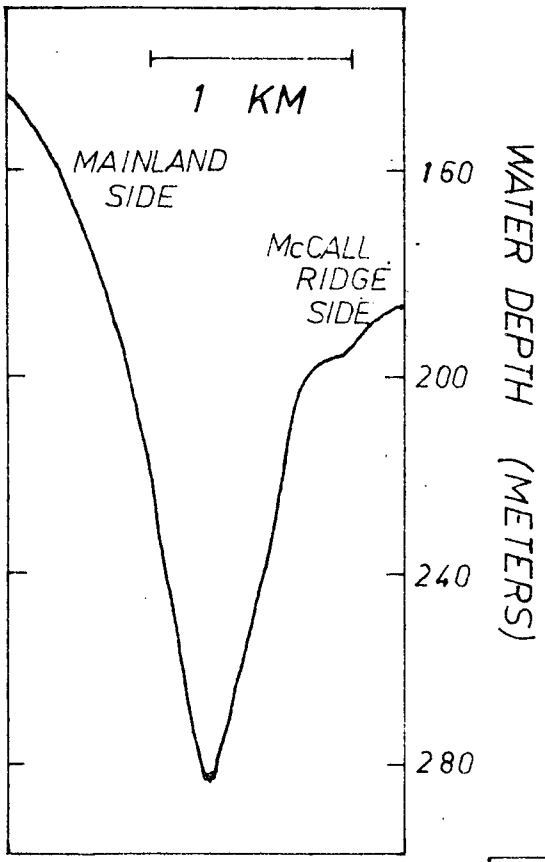
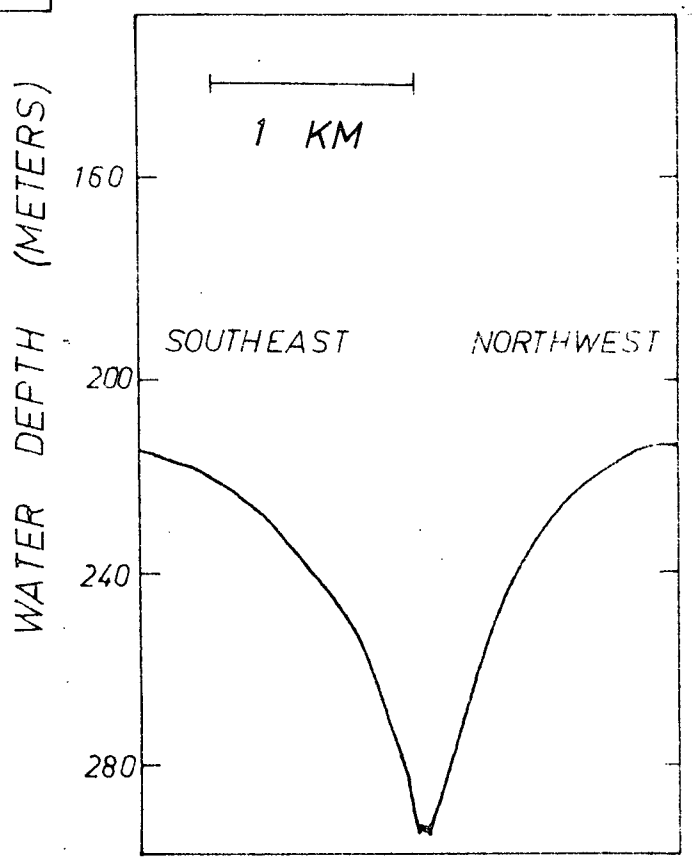


FIGURE 7

Echograms over Jones Deep showing depth and configuration of the hole.



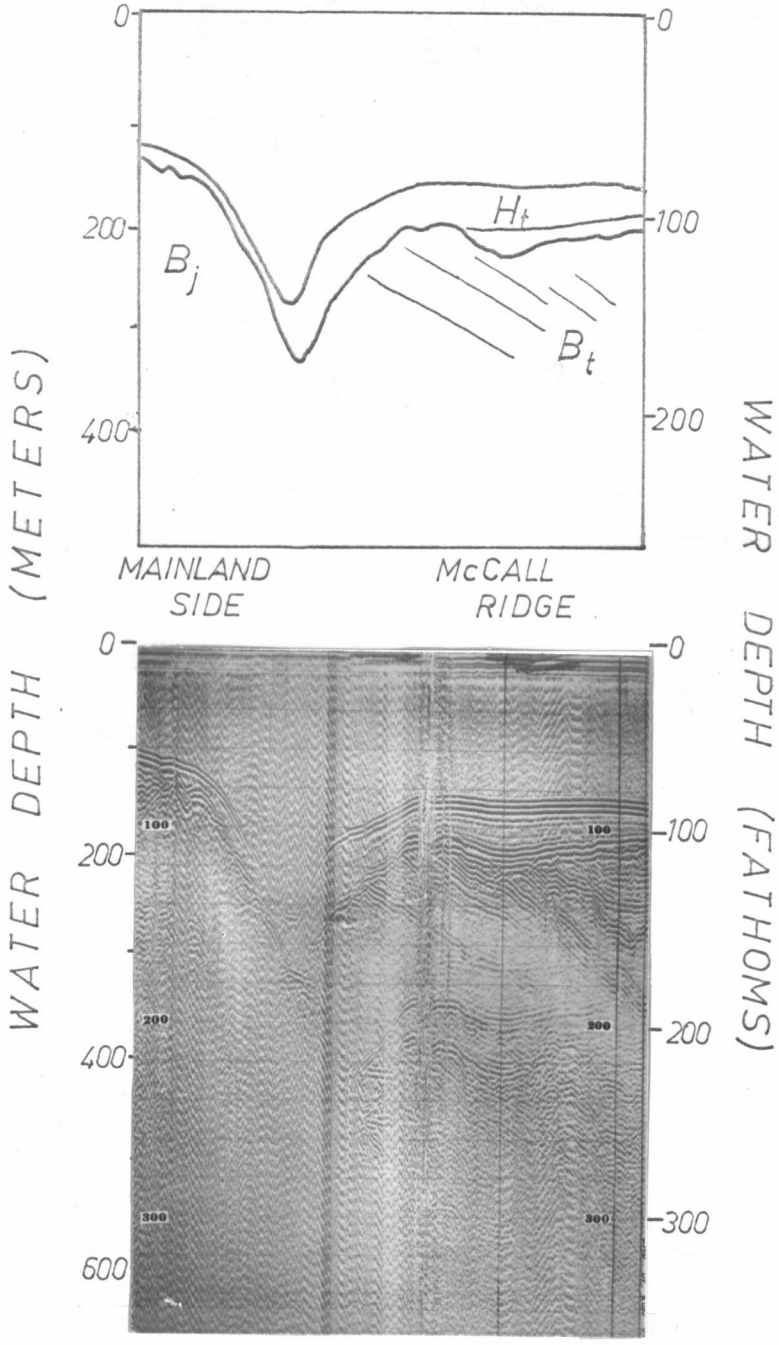


FIGURE 8. Continuous seismic profile over Jones Deep showing depth of unconsolidated sediment surrounding the area.

bulent currents are known. Such currents may, in time, even erode bedrock. A depression at Boundary Pass near Patos Island where tidal currents often reach 1.5 to 2.5 meters per second (3 to 5 knots) may be related to high currents, but over Jones Deep no strong currents are known to exist. Tidal current models suggest tidal streams do move north along the mainland coast (P. Crean, Pacific Oceanographic Group, personal communication) but indicate that currents are very small and decrease to near zero not far to the northwest of Jones Deep.

The possibility that the hole may be a relic of previous topography that has not yet been filled in, is not likely either. Sedimentation rates in the area are high as proven by the depth of sediments on the seismic profiles. Queen Charlotte Trench is covered by at least 180 meters (600 feet) of unconsolidated sediments. McCall Ridge near the hole has a covering of 36 meters (120 feet) and the canyon itself just south of the hole is covered by more than 60 meters (200 feet) of sediment. Figure 8 shows the depth of sediment around the hole. The sides are steep for normal depositional slopes in the area.

A more likely explanation may be that the hole is kept open by a fresh water outlet of a deep aquifer under the canyon. The contact between rock types would be a likely place for such an outlet. Fresh water, being less dense

than salt water, rises toward the sea surface. Circulation set up by the rising water column may be sufficient to prevent sedimentation over the hole and thus permit the present equilibrium to exist. There are no known aquifers of this type reported in the Strait of Georgia, and the Coast Range granitic rock and altered bedrock of Bowen Island and other nearby islands is dense and solid, not normally associated with subterranean water movements. Deep fractures and jointing are common in the rock but some doubt remains that water could be carried from the land to the depth and distance required. Mr. Chris Burton, geologist at Britannia Mine on Howe Sound, a deep mine located in similar rock material, has indicated (personal communication) that ground water circulation in deep areas of the mine is negligible.

None of the proposed explanations are entirely satisfactory from all aspects but, in the author's opinion, the latter still remains the most plausible. Salinity tests for brackish water should be made over the hole.

II. TRINCOMALI TROUGH

A trough, or valley, at the base of the Gulf Island slope connects the southern end of Ballenas Basin to Boundary Basin. It has been named here, Trincomali Trough. The trough is more than 18 kilometers in length but, over most of its course, it rarely exceeds 1.5 kilometers in width at

the 180 meter (100 fathom) contour. The floor slopes from north to south with sill depth at about 187 meters (102 fathoms) at the north end. Cross-sections of the trough are shown in Plates XXIV to XXIX. The configuration changes from north to south, influenced by the underlying bedrock structure of the Gulf Island slope and by more recent deposition. In the north the trough is flat-floored and broad. Bedrock of the island slope, forming the west side, continues under the trough, buried by up to 100 meters (300 feet) of sediments which probably post-date Roberts Swell. The east side was, at an earlier stage, formed by a low Pleistocene ridge. At the time of deposition of the Roberts Swell sediments this was buried to a depth of 30 or more meters. The sediments of Roberts Swell now make up the gentle east rise of the trough as it enters Ballenas Basin.

The central trough area (Plates XXVI to XXVIII) differs in configuration from either end. It is U-shaped with a much thinner sedimentary layer over the bedrock. Roberts Swell sediments form a high eastern wall and the buried Pleistocene ridge is much deeper, well below the bedrock base of the trough and visible only in the gas exploder profile of Plate XXVII.

Toward the south end, shallow bedrock of the island slope extends farther to the east, forcing Trincomali Trough to curve away from the island side. The trough is incised

into the contact between the bedrock and softer Roberts Swell sediments, forming a V-shaped canyon cut to a depth of nearly 90 meters (50 fathoms) below the level of Roberts Swell. No sediments cover the floor of this part of the southern trough but thin layers of sediment, possibly related to Roberts Swell, rest on the island slope to the west.

The present variations in shape of Trincomali Trough over its length are probably relatively recent. At an earlier stage before the southern trough was formed it sloped in the reverse direction. Sediments in Boundary Basin were once more extensive and filled the Basin to a higher level. The trough in the south did not exist at its present depth or shape, but was much shallower. The truncation of Roberts Swell strata prove that the southern trough was cut after deposition of these sediments. In the central section, the conformability of Roberts Swell strata under the side of the trough show that the trough existed here in its present state when Roberts Swell sediments were deposited and has changed but little since. In the north, the trough was much deeper before thick sediments were deposited on its floor. Thus the early slope of the trough was from south to north. Deposition in the north end and erosion in the south changed the slope to its present north to south direction. Whether erosion is continuing or not is unknown. However it is clear that deposition is not occurring in the southern or central part

of the trough. Any sediment transported into those parts must be carried completely through it. Since no Recent sediments are accumulating in Boundary Basin, it is doubtful that sediments are carried to the south unless carried beyond the southern Strait. Movement in the opposite direction from Boundary Basin may occur since large accumulations of sand are found at the north side of Roberts Swell, but these may be from the Swell itself. A program of sediment sampling through Trincomali Trough and around its ends may provide further information on sediment movements and sources in the southern Strait.

III. ANOMALOUS HILLS OF THE FRASER DELTA

Approximately nine kilometers downslope from the mouth of the Fraser River, between depths of 220 to 330 meters (120 to 180 fathoms) an area about 5½ kilometers wide and 9 kilometers long has been described by Mathews and Shepard (1962) as consisting of hummocky topography. Profiles over the area are shown in Plates XX to XXIII.

The hummocks are actually small hills or knolls with about 15 to 30 meters of relief from trough to peak. In cross-section on the continuous seismic profiles they have the appearance of being wavelike undulations in the subsurface delta material rather than random hills or hillocks deposited upon that surface. Although crest-to-crest wave-

length varies, it is normally 600 to 750 meters (2,000 to 2,500 feet). The profile and wavelength downslope is similar to that along slope. The knolls therefore appear to be symmetrical in both directions. It should be pointed out, however, that a sufficiently fine grid of precisely controlled echo sounding profiles has not been established to permit an accurate description and evaluation of the complete hill region. Some hills may have other shapes.

Mathews and Shepard suspect the area to be the remains of former landslides modified by current action and more recent sedimentation. Lying as it does below the area of maximum sedimentation at the river mouth, this suspicion is justified. The upper slopes near the river mouth are relatively steep and cut by deep gullies comparable to landslide gullies off the Mississippi Delta. Slope angles of four degrees or more occur above the 90 meter (50 fathom) contour and locally much steeper slopes may be present (see Mathews and Shepard, 1962, Figure 3). Below 90 meters water depth, but above the anomalous hill area, the slope angle decreases but is still 1.5 degrees. Thus slides from near the river mouth could have slid into the present area of the anomalous hills. This assumption is, at least in part, borne out by the seismic profiles. The additional information not available to Mathews and Shepard is on the thickness of the slide layer and its internal structure.

The seismic profiles show sub-bottom reflectors remain continuous under several hills. These sub-surface reflectors undulate with the crests and troughs of the surface. The depth of movement can thus be traced to at least 75 to 90 meters (250 to 300 feet) under delta sediments before the reflectors weaken and disappear. Beneath the hill areathe next deeper major horizon, about 120 meters (400 feet) below the sediment-water interface, has also buckled. Deeper still, a horizon which has been identified previously as the top of ancient bottomset bedding has also been disturbed. Thus movement may have occurred in a layer several scores of meters thick. Instead of many individual slides, the seismic records imply one massive slide layer which has moved en masse down the slope over a deep glide plane. At the toe of the slide less competent sediments under the hill area have flexured and buckled under the stresses imposed by the mass above. At some places in the hill area the layer appears to have fractured with the upslope end gliding over the lower part (Plate XXI). The appearance on the record of wavelike hills is therefore likely due to flexuring. Some hills under which the internal reflectors are steep and discontinuous are apparently fractured blocks which, in sliding, have rotated backward leaving internal layering exposed. Beyond the slide area, the thickness of the undisturbed sediments consistent with the slide layer measures about 76 meters

(250 feet). If a layer of this original thickness took part in the slide, the effect of crumpling at the toe has thickened the layer by 20 percent or more.

Later sedimentation has filled in and subdued the slide topography. It is interesting to note that troughs higher on the slope have been filled whereas the lower ones have not. If sedimentation has not advanced as far as the lower slopes, then the slide is probably not old. Assuming the sedimentation rate of one foot per year given by Mathews and Shepard (1962, p. 1423) for the area close to the river mouth, the slide could not be less than about sixty years old. However, that rate may be excessive for the distance of the knolls from the river mouth. A better estimate may be made by following the top of the slide layer upslope and measuring the thickness of undisturbed recent delta sediment over it. There, close to the river mouth, the assumed sediment rate will be more accurate. The sediment cover, approximately 49 meters (160 feet) thick, indicates the slide probably occurred some 160 years ago.

IV. SEA LEVEL CHANGES IN THE STRAIT OF GEORGIA

The geological record of the area about the Strait of Georgia, including the Nanaimo Lowland and Fraser Lowland, gives ample evidence of former higher stands of sea level (Armstrong, 1959; Easterbrook, 1963; Fyles, 1963). Lower

stands of sea level are more difficult to show. The evidence may now be covered with water, may also be buried under more recent sedimentation, or may be completely removed by erosion.

The general drowned aspect of the British Columbia coastline and the absence of drift deposits on the outer coast are cited by Peacock (1935) as evidence for former lower relative sea levels. Plates III A, and VII A, of Holland (1964) illustrate the drowned appearance of much of the coastal area. Fyles (1963) admits to the possibility that land stood higher to the sea than it does today since post-Vashon time. In northern Washington, Easterbrook (1963, p. 1481) found trenching in Cherry Point silts and clays below present beach level and presumed that relative sea level had, at one time, been lower. Radio-carbon dating of shells from the upper Cherry Point silt gives an age of 38,000 years B.P., or pre-Olympia Interglaciation. Pleistocene upland areas in the Fraser Valley are separated by deep troughs now covered in several hundred feet of sediment (Armstrong, 1956, p. 5; Easterbrook, 1963, p. 1475). Since the alluvium filled valleys are presently not much above sea level, the troughs may have been eroded during a time when sea level stood hundreds of feet below its present level.

Three ways in which relative sea level changes occur are: 1) by eustatic adjustments of the world ocean levels, 2) by isostatic adjustments of land areas due to Pleistocene

ice loading, and 3) by local or regional tectonic movements. All three have operated in the Strait of Georgia area.

Curray (1961) has shown eustatic changes of sea level over the Late Pleistocene Period reached a maximum at about 20,000 years B.P. Sea level then was 360 feet below its present level. At the same time, the Strait of Georgia area was approaching the end of a non-glacial episode, the Olympia Interglaciation (Armstrong, et al, 1967, p. 324). The area had been relatively ice-free for perhaps 15,000 years or more during which time interglacial sediments had been deposited throughout much of the area. Thus no ice load depressed the land, emergence from previous glacial episodes probably was complete, or nearing completion, and sea level was considerably lower. What tectonic movement was occurring, if any, is not known. The Pacific coastal belt is generally conceded to be a youthful area which, despite the present drowned appearance of the coastline, is undergoing uplift. The net result of the pre-Vashon relative movements may have been that sea level in the area was considerably lower than it is at the present time.

If the stratified sediments presently found under McCall and Halibut Ridges were more extensive in the northern study area as suggested by the remnants of these deposits scattered about that area, it is possible that they were, at least in part, removed by sub-aerial erosion before the onset

of the Vashon ice sheet. However, sub-aerial erosion could not account for the present depths of the Strait of Georgia since these are well below the sill depths of the Strait. Also it is believed that, since the Vashon ice sheet overrode much of the earlier unconsolidated deposits on land, its erosive activity was limited (Johnston, 1923). But glacial erosion may have been much greater in the Strait, the main channel for ice movement.

Evidence for lower sea level in the Strait of Georgia is not generally present in the areas of the Strait covered in this survey. The general level of the elevated terrace area may be indicative of a long stand of the sea near 110 fathoms. West of Thormanby Islands it stands near 90 fathoms (Plate II) and is cut in what are probably pre-Vashon Pleistocene sediments. However, no consistent terrace depths are observed throughout the Strait area. It is possible that evidence could be found closer to shore where the ship could not go. Hydrographic charts do show reefs and shoals at various depths, usually less than five fathoms, along the coasts and near the islands.

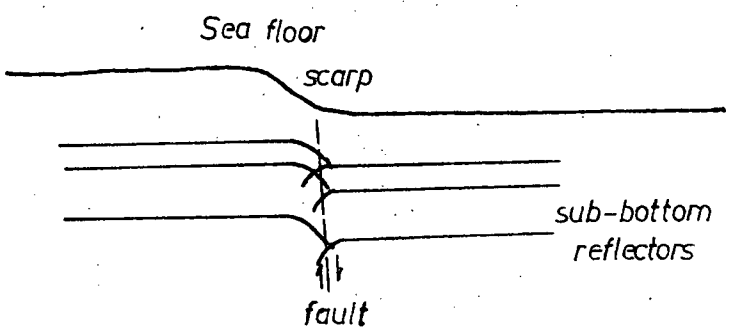
V. FAULTING AND TECTONIC MOVEMENTS

Identification of faults on seismic records is often difficult and sometimes tenuous. Several criteria for fault identification from conventional seismic records are noted

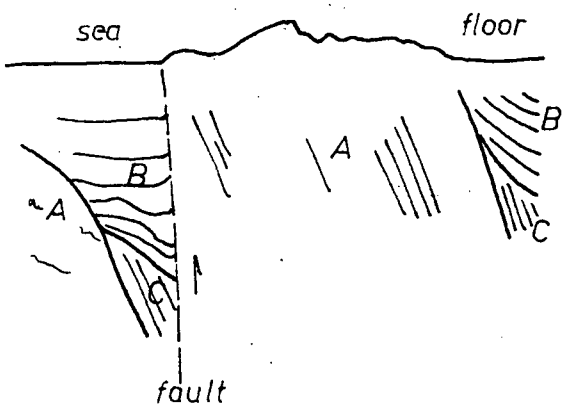
in the literature (Campbell, 1965). The density of information provided by the continuous seismic method make some criteria more useful than others. Correlation of seismic events across a vertical displacement, or even a zone of disruption, may indicate faulting, especially if the sea bottom too, has been displaced and forms a scarp. A line drawing of a fault from another area identified in this way is given in Figure 9(a). Horizontal as well as vertical displacement on a fault may be recognized by a repeated pattern of reflections. Such a fault is shown in the line drawing of Figure 9(b) from an actual seismic record. Local disruptions of stratification in which no correlation can be made across the disrupted area may be indicative of faulting but other causes can give similar effects on the record. Further proof is then needed to make positive identification.

Also indicative of faulting, but not proof, is a vertical displacement of the sea floor when no reflector, or little penetration, is present. Because of exaggeration usually present on CSP records, angles appear much steeper than they actually are, and these steep-looking slopes can easily be confused as a fault scarp. Topographical expression as a fault criteria must therefore be used with caution.

No faults have been well identified from the seismic profiles over the study area. Nevertheless it is suspected that faults do occur. In the Nanaimo Basin area major faults



(a) Normal fault in plane reflectors.



(b) Repeated pattern of seismic reflectors indicates fault with lateral movement.

FIGURE 9. Fault criteria appearing on CSP records.

have been reported by Buckham (1947), Clapp (1912, 1914), Muller (personal communication), and others. Bell (1967) postulates a fault extending along the mainland side of the Strait of Georgia.

On the basis of topographic character alone, the en echelon step-like ridges of the Gulf Island slope could be assumed to have fault origins. The seismic profiles, by penetrating the sediment filled troughs between the ridges and exposing their bases as well as other buried crests, further this impression but by no means prove it. The ends of bedding planes are truncated on the steeper island side of the ridge. Thus some ridges at least are cuesta-like. Like cuestas, they may have faults parallelling them. Earthquake activity is common in the Gulf Islands (Milne, 1963) near this area.

The most westerly profile beyond the end of Ballenas Basin, Plate II, shows what can be interpreted as fault blocks southeast of Sangster Island. The blocks are tilted toward Lasqueti Island. Bolson-like features formed by the tilted blocks have been filled by later sediments or till. Muller (personal communication) has recorded the presence of large tilted fault blocks on Texada Island nearby. White and Savage (1965) note the possibility of major faulting not far to the north of this area.

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Other steep-sided slopes of islands and ridges, such as Ballenas Islands, may have fault origins, but no definite seismic information is available in the way of proof.

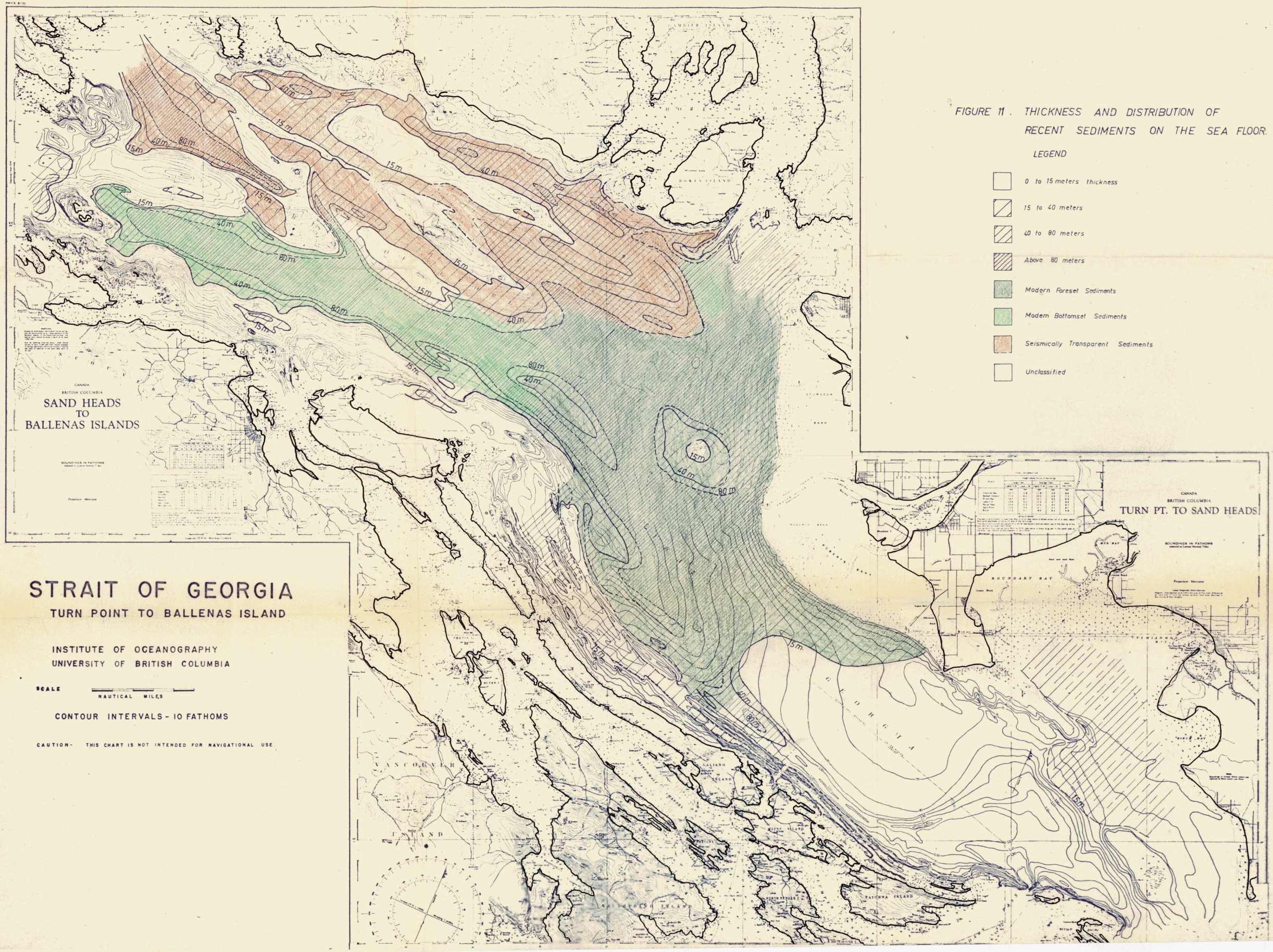


FIGURE 11. THICKNESS AND DISTRIBUTION OF RECENT SEDIMENTS ON THE SEA FLOOR

LEGEND

- 0 to 15 meters thickness
- 15 to 40 meters
- 40 to 80 meters
- Above 80 meters
- Modern Foreset Sediments
- Modern Bottomset Sediments
- Seismically Transparent Sediments
- Unclassified

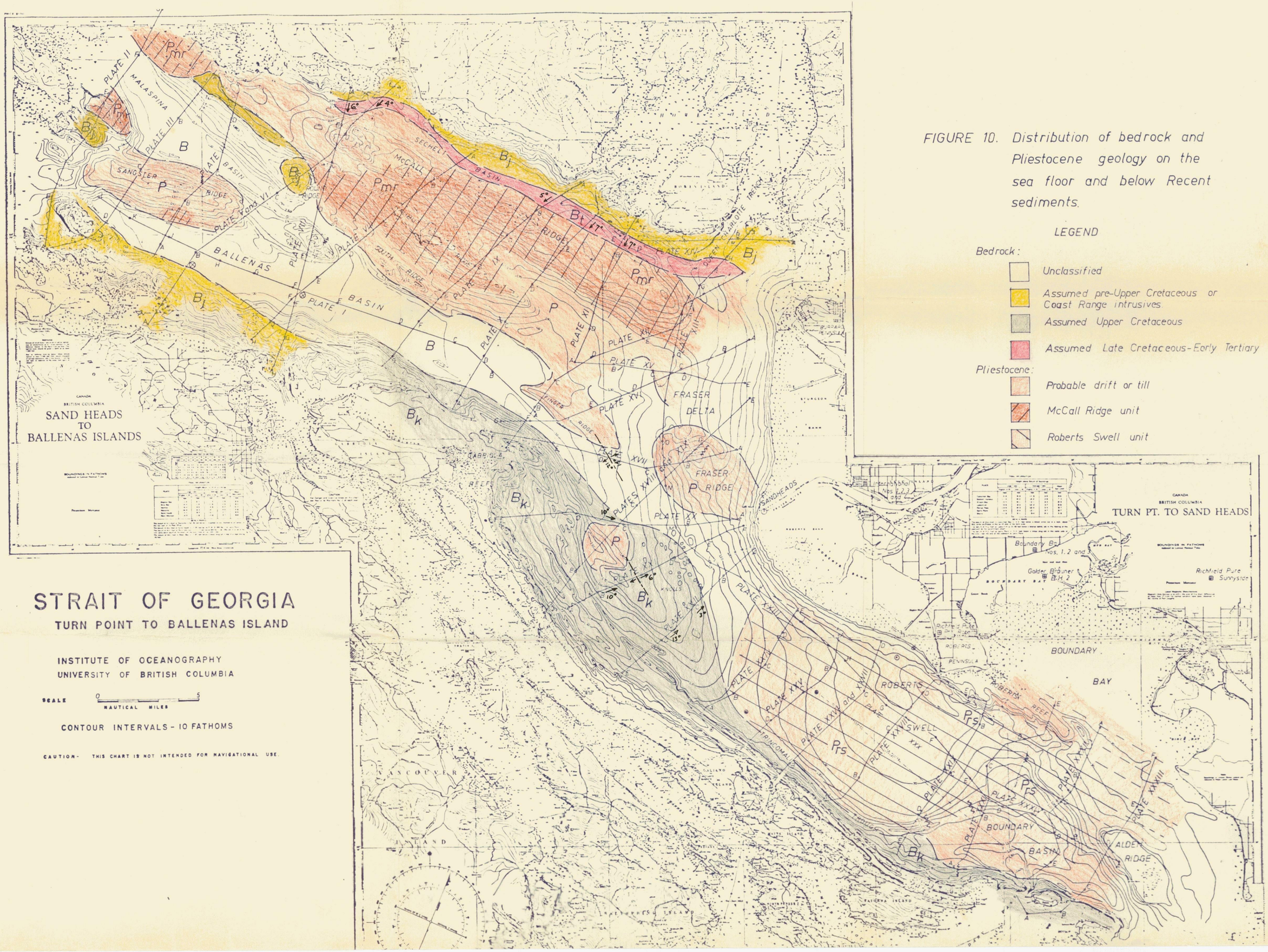


FIGURE 10. Distribution of bedrock and Pliestocene geology on the sea floor and below Recent sediments.

CHAPTER VI

CONCLUSIONS

I. USE OF THE CONTINUOUS REFLECTION PROFILER IN BRITISH COLUMBIAN COASTAL WATERS

The continuous seismic profiler is an invaluable tool for the geophysicist or geologist interested in studying the upper crustal structure of the sea floor off British Columbia. While excellent penetration can be obtained in most soft unconsolidated or semi-consolidated sediments, bedrock penetration depends upon the character of the rock, the degree of smoothness, and the slope of its upper boundary. Rough and irregular boundaries scatter sound energy and reflect hyperbolic patterns which obscure structure below the boundary. However large scale features on the bedrock surface can often be recognized. In areas where sound penetration gives meaningful results in bedrock, much useful information such as direction and degree of structural dip, synclinal and anticlinal axes, and perhaps evidence of faulting, can be obtained. Because sound velocity is commonly much greater in indurated rock, apparent low penetration in the time domain may be great in terms of actual depth.

Continuous seismic profiling is extremely useful where a mantle of till and relict Pleistocene or Recent sedi-

ment obscures the earlier surface and its nature to most other methods of study.

From record character it is possible to distinguish various sedimentary facies. Soft unconsolidated sediments can be recognized overlying areas of semi-consolidated sediments or bedrock. Where deep echoes are weak, it may be difficult to separate bedrock from semi-consolidated or other sediments but, in many cases, the top of bedrock can be traced from other areas where it is more definite and easily recognized. In fact, one great advantage of continuous seismic profiling is its ability to locate outcrop areas where sampling can be used to maximum advantage while the extent of the sampled layer can be traced much farther under the overburden.

Like all geophysical methods, the CSP method is most useful in association with other techniques or measurements which provide additional information. Accurate structural information is available only if velocities are obtained, for example, by refraction or wide-angle reflection methods using CSP equipment with sonobuoys (LePichon, et al, 1968; Houtz, et al, 1968). Gravity and magnetic measurements provide further useful information on bedrock structures. Sampling of the sea bottom is indispensible to a CSP study in order to determine the nature of the material underlying the sea floor.

II. SUMMARY OF GEOLOGY AND STRUCTURE UNDER THE STRAIT OF GEORGIA

Bedrock Structures

In the study area of the Strait of Georgia, sedimentary areas have been outlined down to the bedrock surface, and in some areas, structural features in the bedrock itself have been delineated. Under the Strait bedrock falls into two general categories: seismically opaque, and seismically stratified. Unfortunately, many rock types such as Coast Range granodiorites, volcanic and basic rock, pre-Upper Cretaceous meta-sediments and some bedrock of Upper Cretaceous age are apparently opaque. These rocks cannot be distinguished one from the other in this area by their seismic character. Where tentatively identified, the identification is due to proximity of shoreline outcrops to the recording site. The distribution of bedrock and Pleistocene deposits under the sea floor of the Strait is shown in Figure 10, back pocket.

Bedrock showing seismic stratification is present under several parts of the Strait, the two main areas being along the Gulf Island slope south of Gabriola Reefs (marked B_k), and in the McCall Ridge and Sechelt Basin area along the northern mainland (marked B_t).

Along the northern mainland, between the north and south ends of McCall Ridge, the strata dip to the southwest at angles of approximately 5 to 10 degrees or 90 to 180 meters

per kilometer. The top of bedding is found about 245 meters (800 feet) below sea level at a distance of two to three kilometers offshore. Onshore, Coast Range granodiorites dominate the coastline with a few older meta-sedimentary outliers occurring near Howe Sound. The mainland slope between the shoreline and truncated strata offshore is doubtless composed of granodiorite or the older meta-sedimentary rocks near Bowen Island. The truncated strata overlies these rocks and dips under thick Pleistocene deposits farther offshore. The stratigraphic position and even the dip is similar to that of Late Cretaceous and Early Tertiary rocks resting unconformably on granitic rocks of the Coast Mountains and underlying Burrard Peninsula at Vancouver, not far distant. On this basis the section of stratified bedrock near the mainland coast has been assigned to the Late Cretaceous-Early Tertiary Period. What happens to these rocks farther under Georgia Strait has not been ascertained. The strata disappear under McCall or Halibut Ridges and, except for minor stratification appearing along the axial profile of Ballenas Basin, (Plate I) no stratified bedrock occurs to the opposing Vancouver Island slope.

Bedrock along the coast of Vancouver Island opposite the suspected Late Cretaceous-Early Tertiary area ranges from Upper Cretaceous at Gabriola Island to pre-Upper Cretaceous sedimentary and volcanic rocks north of Nanaimo, all seismi-

cally opaque. If the dip of the Tertiary rocks along the mainland side remained at 10 degrees into Ballenas Basin, a thickness of greater than 3,400 meters (11,000 feet) should be present at the base of the Vancouver Island slope. It is doubtful that such is the case but the presence of even some stratification in the bedrock under Ballenas Basin is suggestive of the presence of a rock type more characteristic of the northeast side of the Strait. Thus, it is possible that the Vancouver Island slope marks a fault scarp with the basin side down. Tertiary rocks may be in fault contact with older units on Vancouver Island. If such a fault should exist, it is most certainly a major one. Lack of good bedrock reflectors under the basin in the cross-profiles prevents more positive interpretation.

Stratification along the Gulf Island slope does not appear to be continuous with the stratified unit along the northern mainland, but may be related. In this area the strata dip to the east and can be followed a short way under the Fraser Delta area where the dip decreases and the strata are deformed into gentle synclinal and anticlinal structures. This stratified bedrock is, by its position and structural relationship to the Gulf Island slope, associated with the Upper Cretaceous formations on the Gulf Islands. However, because the strata obviously remain near the surface and may even rise under Fraser Ridge, a relationship with the Late

Cretaceous-Early Tertiary bedrock under the Fraser Lowland area is possible (Crickmay and Pocock, 1963). Therefore, unless a large fault runs down the Strait under the Fraser Delta as postulated by Bell (1967), the lower part at least of the Burrard Formation on the mainland and the Upper Cretaceous Gabriola Formation on the Gulf Islands may be contiguous. Muller (Geological Survey of Canada, personal communication) believes the Gabriola Formation may be, at least in part, Tertiary.

Under the southern Gulf Island slope bedrock has a steeper northeasterly dip than bedrock under the island slope to the northwest. The dip continues steep to the point where bedrock disappears under Roberts Swell. Beyond there, only tops of bedrock are seen in some profiles.

Small areas of bedrock stratification appear under Boundary Basin but because they are not continuous with the island slope or other areas where the geology is known, their relationships are not clear. From the depth below sea level it is thought that they may be part of Tertiary bedrock as found in boreholes in the Fraser Lowlands, but once again, the proximity of island slope bedrock of probable Upper Cretaceous age make even this uncertain. Deformation appears to be more intense in the bedrock under Boundary Basin than elsewhere on the island slope. In other areas of the Strait,

the top of bedrock is recorded but little or no penetration is achieved.

Although evidence is not clear, the profiles suggest that the Strait of Georgia area may have undergone a flexure with upwarping of the Coast Ranges and depression of the basin axis close to the Vancouver Island side. The Strait of Georgia is consistently deeper on this side. The island side may then have fractured along the Vancouver Island slope. If the system of ridges along the Gulf Island slope is fault controlled, their convergence in the south indicates a more concentrated fault zone in that area.

Before the deposition of Pleistocene sediments, the Strait of Georgia was a long valley floored by bedrock which sloped gently off the Coast Range Mountains toward Vancouver Island. The valley extended from north of the study area to the San Juan Archipelago and probably included the re-entrant now occupied by the Fraser Lowlands, much of which was developed during the Pleistocene. Relief on the valley floor was likely considerable as evidenced by the existence of breached anticlines and bedrock ridges now buried in more recent sediments.

Pleistocene Geology

During the Pleistocene large deposits of till and interglacial sediments were left in the Strait. The distribution of these deposits is shown in Figure 10. In the northern

study area they tend to range along the mainland side of the Strait whereas in the area south of the Fraser Delta, Pleistocene deposits occur across the Strait. At least two areas where extensive deposits can be traced as a single unit have been found. McCall and Halibut Ridges, although underlain by bedrock, consist mainly of thick sections of Pleistocene. These undersea ridges are comparable in magnitude and structure to the Pleistocene ridges of Burrard Peninsula and Burnaby Mountain at Vancouver. The McCall Ridge unit is recognized from near Point Grey to at least as far as the north end of North Thormanby Island with perhaps a short break south of South Thormanby Island. It may be as much as 560 meters (1,800 feet) thick in places. Bedding in the unit is flat or exhibits only low dips. Truncation of the beds suggests erosion has occurred. The unit may have covered much more of the Strait than the area presently occupied.

South of the Fraser Delta, the second large sedimentary deposit, the Roberts Swell unit of probable Late Pleistocene age, covers a large part of the southern Strait of Georgia including Roberts Swell, the mainland shelf, most of Boundary Basin and at least the northwestern part of Alden Ridge. This unit extends close to the base of the Gulf Island slope. Older Pleistocene deposits occur below it along Trincomali Trough and near Roberts Peninsula. Under Boundary Basin truncation of this unit around the perimeter

of the basin indicate that it once existed to a greater depth, but has since been eroded in Late Pleistocene to Recent time. As the area is presently one of non-deposition, and relatively high velocity tidal currents occur, erosion may still be active.

Other Pleistocene deposits which occur are not continuous over a large area or cannot be correlated from profile to profile. Many ridges in the Strait are topped by Pleistocene sediments. However, the bases of at least some ridges are bedrock. Only Round Ridge is completely bedrock, possibly a boss of Coast Range granitic material. Sangster Ridge is, if not wholly Pleistocene, at least in part composed of sediments of that age. Other ridges such as Finger Ridge, South Ridge and Fraser Ridge, also appear to be capped by Pleistocene sediments. No Pleistocene is found along the Vancouver Island slope.

Glacial erosion may have been particularly powerful on the Vancouver Island side if the present suspended valleys in the bedrock of the slope are actually 'hanging' valleys. The oversteepened lower slopes of Ballenas Basin also point to heavy glacial erosion. It is possible that most of the volume of ice movement to the southeast through the Strait was concentrated on the Vancouver Island side through Ballenas Basin. No Pleistocene deposits are recognized in the basin. Basin fill is Recent.

The profile of Malaspina Basin is V-shaped with sides less steeply inclined than in Ballenas Basin indicating less glacial erosion and perhaps less ice movement.

The Pleistocene deposits are probably till, drift and inter-glacial in origin. On the continuous seismic records, unstratified, irregularly stratified or chaotic reflection characteristics, often broken by many unconformities, are thought to be evidence of till or drift materials, whereas stratified material is inter-glacial or glacial. Several glacial episodes occurred in the Strait but, because of the rapid lateral as well as vertical changes in the Pleistocene sediments, the number of separate glacial occurances cannot be determined from these records. If the McCall Ridge unit is accepted as a single inter-glacial formation, then it seems likely that it separates at least two glacial episodes above and below it. However, more than one glacial episode may have occurred prior to, or since, deposition of the McCall Ridge unit.

Recent Sediments

Recent sediments in the Strait have as their principal source, the Fraser River. Sediment movement is mainly to the north from the delta and little, if any, sedimentation is presently occurring in the Strait south of the delta. Distribution and thickness of Recent sediments is shown in

Figure 11, back pocket. The modern Fraser Delta, which is probably no more than 10,000 years in age, has extended across the Strait to the Gulf Island slope, overlying the north side of Roberts Swell, but not occurring over the top of the Swell. Little Recent sedimentation is found in Boundary Basin to the south. No southward sediment transport appears to be occurring except perhaps that which continues beyond the study area. Sediments brought to the Strait by the Fraser River are now deposited on the delta front or distributed to the north and northwest mainly into the basin areas. The main sediment catchment area is Ballenas Basin where sediments exceed 180 meters (600 feet) in average thickness even at distances of 50 kilometers from the Fraser River. Sechelt Basin and Malaspina Basin also receive considerable quantities of Recent sediments.

Recent sediments also cover ridge areas along the mainland north of Burrard Inlet and are thick near the Fraser Delta, but they thin rapidly and become spotty as the distance from the delta increases. In many areas, slopes into the basins are free of Recent sediment cover even though their angle is low enough to support such cover. Since no slump or slide deposits are found under the sides it is suspected that currents are responsible for sweeping the sidewalls clear.

The transparent seismic quality of sediments over the ridge tops and in some basin areas is in contrast with stratification appearing in sediments in Ballenas Basin. The stratification in the latter basin may be due in part to turbidity currents originating on the Fraser Delta front.

A major change in Recent sedimentation is marked by a strong reflecting horizon in thick sediments over Ballenas Basin and other smaller and local basin areas. The sediments below the reflector are suspected to be ancient pre-delta beds from a time before the delta foreset beds extended over-top McCall-Fraser Ridges. The Fraser River may have entered the Strait farther to the south and east than it presently does. As the delta built seaward toward its present site, sediments began to spill over the sides of the ridges perhaps causing the change in sedimentation which created the seismic discontinuity noted over a wide area.

Few Recent sediments occur along the Vancouver Island slope except locally in pockets or depressions in the bedrock. The floor of the hanging valley north of Gabriola Island is only thinly covered.

REFERENCES

Armstrong, J.E., 1956: Surficial geology of the Vancouver area, British Columbia. Geol. Survey Canada, Paper 55-4, 16 pp.

Armstrong, J.E., 1957: Surficial geology of New Westminster map-area, British Columbia. Geol. Survey Canada, Paper 57-5, 25 pp.

Armstrong, J.E., 1960a: Geology of the coast mountains, North Vancouver, British Columbia. "Guidebook for Geological Field Trips in Southwestern British Columbia." Geological Discussion Club, Vancouver, 15-25.

Armstrong, J.E., 1960b: Surficial geology of Sumas map-area, B.C. Geol. Survey Canada, Paper 59-9, 27 pp.

Armstrong, J.E., and W.L. Brown, 1954: Late Wisconsin marine drift and associated sediments of the lower Fraser Valley, British Columbia. Geol. Soc. America Bull., 65, 349-364.

Armstrong, J.E., D.R. Crandell, D.J. Easterbrook and J.B. Noble, 1965: Late Pleistocene stratigraphy and chronology in southwestern British Columbia and northwestern Washington. Geol. Soc. America Bull., 76, 321-330.

Bacon, W.R., 1957: Geology of lower Jervis Inlet, British Columbia. British Columbia Dept. of Mines, Bull. 39, 45 pp.

Bancroft, J.A., 1913: Geology of the coast and islands between the Strait of Georgia and Queen Charlotte Sound, British Columbia. Geol. Survey Canada, Mem. 23, 146 pp.

Beckmen, W.C., A.C. Roberts and B. Luskin, 1959: Sub-bottom depth recorder. Geophysics, 24, 749-760.

Bell, G.L., 1967: Big potential oil basins offshore from west coast. Oilweek, June 27, 1967, p. 48.

Bennett, Lee C. Jr., and S.M. Savin, 1963: The natural history of the Hardangerfjord. 6. Studies of the sediments of parts of the Ytre Samlafjord with the continuous seismic profiler. Sarsia, 14, 79-94.

Backus, M.M., 1959: Water reverberations--their nature and elimination. Geophysics, 24, 233-261.

Bretz, J.H., 1913: Glaciation of the Puget Sound region. Wash. Geol. Survey Bull., 8, 244 pp.

Buckham, A.F., 1947: The Nanaimo coal field. Trans Canada Inst. Mining Met., 50, 460-472.

Burwash, E.M.J., 1918: Geology of Vancouver and vicinity. University of Chicago Press, Chicago, Ill., 106 pp.

Campbell, Francis F., 1965: Fault criteria. Geophysics, 30, 976-997.

Clapp, C.H., 1912: Southern Vancouver Island. Geol. Survey Canada, Mem. 13, 208 pp.

Clapp, C.H., 1913: Geology of the Victoria and Sanich map area, Vancouver Island, British Columbia. Geol. Survey Canada, Mem. 36, 208 pp.

Clapp, C.H., 1914: Geology of the Nanaimo map area. Geol. Survey Canada, Mem. 51, 133 pp.

Clark, S.P. Jr., 1966: Handbook of Physical Constants. Geol. Soc. America, Mem. 97.

Cockbain, A.E., 1963a: Submarine topography and sediment thickness in the southern Strait of Georgia. Inst. Oceanography, Univ. of British Columbia, Manuscript Report No. 14, 8 pp.

Cockbain, A.E., 1963b: Distribution of foraminifera in Juan de Fuca and Georgia Straits, British Columbia, Canada. Contributions from the Cushman Foundation for Foraminiferal Research, 14, Part 2, 37-57.

Cone, R.A., N.S. Neidell and K.E. Kenyon, 1963: The natural history of the Hardangerfjord. 5. Studies of the deep water sediments with the continuous seismic profiler. Sarsia, 14, 61-78.

Crickmay, C.H., and S.A.J. Pocock, 1963: Cretaceous of Vancouver, British Columbia, Canada. Amer. Assoc. Pet. Geol., Bull. 47, 1928-1942.

Curray, J.R., 1961: Late quaternary sea-level: A discussion. Geol. Soc. America Bull., 72, 1707-1712.

Curray, J.R., and D.G. Moore, 1964: Pleistocene deltaic progradation of continental terrace, Costa de Nayarit, Mexico. In "Marine Geology of the Gulf of California." Amer. Assoc. Pet. Geol., Mem. 3, 193-215.

Delingher, P., M. Gemperle, R.W. Couch and D.W. Hood, 1966: Free air gravity anomalies along the inside passage of British Columbia and Alaska. Jour. Geophys. Research, 71, 6011-6015.

Easterbrook, D.J., 1963: Late Pleistocene glacial events and relative sea level changes in northern Puget Lowland, Washington. Geol. Soc. America. Bull., 74, 1465-1484.

Ewing, J.I., and G.B. Tirey, 1961: Seismic profiler. Jour. Geophys. Research, 66, 2917-2927.

Fyles, J.G., 1963: Surficial geology of Horne Lake and Parksville map-areas, Vancouver Island, British Columbia. Geol. Survey Canada, Mem. 318, 142 pp.

Grant, F.S., and G.F. West, 1965: Interpretation theory in applied geophysics. International Series in Earth Sciences, McGraw-Hill Book Company, New York. 583 pp.

Halstead, E.C., and A. Treichel, 1966: Groundwater resources of the coastal lowlands and adjacent islands, Nanoose Bay to Campbell River, east coast, Vancouver Island. Geol. Survey Canada Bull., 144, 42 pp.

Hamilton, E.L., 1956: Low sound velocities in high-porosity sediments. Jour. Acoust. Soc. America, 38, 16-19.

Hamilton, E.L., 1959: Thickness and consolidation of deep-sea sediments. Geol. Soc. America Bull., 70, 1399-1424.

Hamilton, E.L., 1965: Sound speed and related physical properties of sediments from experimental Mohole (Guadaloupe site). Geophysics, 30, 257-261.

Hamilton, E.L., 1967: Marine geology of abyssal plains in the Gulf of Alaska. Jour. Geophys. Research, 72, 4189-4223.

Hamilton, E.L., G. Shumway, H.W. Menard and C.J. Shiper, 1956: Acoustic and other physical properties of shallow water sediments off San Diego. Jour. Acoust. Soc. America, 28, 1-15.

Hersey, J.B., 1963: Continuous reflection profiling. In "The Sea", Vol. 3, The Earth Beneath the Sea, Interscience Publishers, New York, 47-84.

Hersey, J.B., H.E. Edgerton, S.O. Raymond and G. Hayward, 1961: Pingers and thumpers advance deep-sea exploration. Jour. Instrument Soc. America, 8, 72-77.

Holland, S.S., 1964: Landforms of British Columbia. A physiographic outline. British Columbia Dept. of Mines and Petroleum Resources, Bull. 48, 138 pp.

Holtedahl, H., 1965: Recent turbidites in the Hardanger fiord, Norway. In "Submarine Geology and Geophysics." Colston Research Society. Vol. XVII, Bristol, England. 107-140.

Hopkins, W.S., 1966: Palynology of tertiary rocks of Whatcom Basin, British Columbia. Unpublished Ph.D. Thesis. Dept. of Geol., Univ. of B.C., 184 pp.

Houtz, R., J. Ewing and X. LePichon, 1968: Velocity of deep-sea sediments from sonobuoy data. Jour. Geophys. Research, 73, 2615-2641.

Johnston, W.A., 1921: Sedimentation of the Fraser river delta. Geol. Survey Canada, Mem. 125, 46 pp.

Johnston, W.A., 1923: Geology of the Fraser Delta map-area. Geol. Survey Canada, Mem. 135, 87 pp.

King, L.H., 1965: Use of a conventional echo-sounder and textual analysis in delineating sedimentary facies. Bedford Inst. of Oceanography, Report 65-14, 27 pp.

Knott, S.T., 1962: Use of precision graphic recorders in oceanography. Marine Sciences Inst., 1, 251-262.

Krause, D.C., 1962: Interpretation of echo-sounder profiles. International Hydrographic Review, 39, 65-123.

LePichon, X., J. Ewing and R.E. Houtz, 1968: Deep sea sediment velocity determination made while reflection profiling. Jour. Geophys. Research, 73, 2597-2614.

LeRoy, O.E., 1908: Preliminary report on a portion of the main coast of British Columbia and adjacent islands included in the New Westminster and Nanaimo districts. Geol. Survey Canada, 56 pp.

Lindsay, R.B., 1960: Mechanical radiation. International Series in Pure and Applied Physics. McGraw-Hill Book Co., Toronto, 415 pp.

Luskin, B., B.C. Heezen, M. Ewing and M. Landisman, 1954: Precision measurement of ocean depths. Deep-Sea Research, 1, 131-140.

Mathews, W.H., 1958: Geology of the Mount Garibaldi map-area, southwestern British Columbia. Geol. Soc. America Bull., 69, 161-198.

Mathews, W.H. and F.P. Shepard, 1962: Sedimentation of the Fraser River delta, British Columbia. Amer. Assoc. Pet. Geol., Bull. 46, 1416-1443.

Mayers, I.R., 1968: Analysis of the form and origin of the Fraser River Delta's subaqueous slump deposits. Unpublished M.Sc. Thesis. Dept. of Geophysics, Univ. of British Columbia.

McConnell, R.G., 1914: Texada Island, B.C. Geol. Survey Canada, Mem. 58, 112 pp.

McLure, C.D., H.F. Nelson and W.B. Huckaby, 1958: Marine sonoprobe system--new tool for geologic mapping. Amer. Assoc. Pet. Geol., Bull. 42, 701-716.

Milne, W.G., 1963: Seismicity of western Canada. Boletin Bibleografico de Geofisica y Oceanografica Americanas, 3, 17-40.

Milne, W.G., and W.R.H. White, 1960: A seismic survey in the vicinity of Vancouver Island, B.C. Pub. Dom. Obs., 24, 145-154.

Moore, D.G., 1960: Acoustic reflection studies of the continental shelf and slope off southern California. Geol. Soc. America Bull., 71, 1121-1136.

Nafe, J.E., and C.L. Drake, 1957: Variation with depth in shallow and deep water marine sediments of porosity, density and the velocities of compressional and shear waves. Geophysics, 22, 523-552.

Newcomb, R.C., J.E. Sceva and O. Stromme, 1949: Ground water resources of western Whatcom Country. U.S. Geol. Survey, open file report, 135 pp.

Officer, C.B., 1955: A deep sea seismic reflection profile. Geophysics, 20, 270-282.

Officer, C.B., 1958: Introduction to the theory of sound transmission. McGraw-Hill Book Co., 458 pp.

Ostericher, Charles Jr., 1965: Bottom and sub-bottom investigation of Penobscot Bay, Maine, 1959. U.S. Navy Oceanographic Office, Tech. Report TR-173.

Paitson, L., C.H. Savit, D.M. Blue and W.A. Knox, 1964: Reflection survey at Barracuda fault. Geophysics, 29, 941-950.

Peacock, M.A., 1935: Fiordland of British Columbia. Geol. Soc. America Bull., 46, 633-696.

Phemister, T.C., 1945: The coast range batholoth near Vancouver, B.C. Quart. Jour., Geol. Soc. London, 101, pts. 1 and 2, 37-88.

Pickard, G.L., 1956: Surface and bottom currents in the Strait of Georgia. Jour. Fisheries Research Board Canada, 13, 581-590.

Roddick, J.A., 1965: Vancouver North, Coquitlam and Pitt Lake map-area, B.C. Geol. Survey Canada, Mem. 335.

Savit, C.H., W.A. Knox, D.M. Blue and L. Paitson, 1964: Reflection and velocity profiles at the Outer Ridge, Puerto Rico. Jour. Geophys. Research, 69, 701-719.

Shepard, F.P., 1963: Submarine geology. 2nd Ed. Harper's Geoscience Series. Harper and Row, Publishers New York. 557 pp.

Shor, G.G. Jr., D.G. Moore and W.B. Huckaby, 1963: Deep sea tests of a new non-explosive reflection profiler. Jour. Geophys. Research, 68, 1567-1571.

Smith, W.O., 1958: Recent underwater surveys using low frequency sound to locate shallow bedrock. Geol. Soc. America Bull., 69, 69-97.

Smith, W.O., C.B. Cumming and J.E. Upson, 1954: Mapping bedrock with low frequency sound. Geol. Soc. America Bull., 65, (Abstract), 1307.

Stacey, R.A., L.E. Stephens, R.V. Cooper and B.G. Brule, 1969: Gravity measurements in British Columbia. Gravity Map Series, Map No. 88. Dom. Obs., Ottawa.

Tiffin, D.L., and J.W. Murray, 1966: Mapping offshore shelf with continuous seismic. Oilweek, 17, No. 38, 49-51.

Tseng, K.H., 1968: A new model for the crust in the vicinity of Vancouver Island, B.C. Unpublished M.Sc. thesis. Dept. of Geophysics, Univ. of British Columbia.

Usher, J.L., 1952: Ammonite faunas of the Upper Cretaceous rocks of Vancouver Island, British Columbia. Geol. Survey Canada Bull. 21, 182 pp.

Walcott, R.I., 1967: Bouguer anomoly map of southwestern British Columbia. Inst. of Earth Sciences, Scientific Report No. 15, Univ. of British Columbia.

Waldichuk, M., 1952: Oceanography of the Strait of Georgia. 1. Salinity distribution. Progress Reports of the Pacific Coast Stations, Fisheries Research Board Canada, 93, 26-29.

Waldichuk, M., 1953: Oceanography of the Strait of Georgia. 2. Temperature distribution. Progress Reports of the Pacific Coast Stations, Fisheries Research Board Canada, 94, 19-23.

Waldichuk, M., 1954: Oceanography of the Strait of Georgia. 3. Character of the bottom. Progress Reports of the Pacific Coast Stations, Fisheries Research Board Canada, 95, 59-63.

Waldichuk, M., and S. Tabata, 1955: Oceanography of the Strait of Georgia V. Surface Currents. Progress Reports of the Pacific Coast Stations, Fisheries Research Board Canada, 104, 30-33.

Waldichuk, M., 1956: Oceanography of the Strait of Georgia. 6. Fresh water budget. Progress Reports of the Pacific Coast Stations, Fisheries Research Board Canada, 107, 24-27.

Waldichuk, M., 1957: Physical oceanography of the Strait of Georgia, British Columbia. Jour. Fish. Research Board Canada, 14, 321-486.

White, W.H., 1959: Cordilleran Tectonics in British Columbia. Amer. Assoc. Pet. Geol., Bull. 43, 60-100.

White, W.R.H., and J.C. Savage, 1965: A seismic refraction and gravity study of the earth's crust in British Columbia. Seis. Soc. Amer., Bull. 55, 463-486.

White, W.R.H., 1962: The structure of the earth's crust in the vicinity of Vancouver Island as ascertained by seismic and gravity observations. Ph.D. thesis. Dept. of Geophysics, Univ. of British Columbia. 97 pp.

APPENDIX

SOUND VELOCITIES IN WATER, SEDIMENTS AND ROCKS OF THE THESIS AREA

In order to convert the time data recorded on the seismic profiles to depth data, the velocity of sound waves over the travel path of the signal must be accurately known. This knowledge permits the correct determination, not only of depth, but of thickness, shape, slope and dip as well. It is possible that reconstructed structures may differ significantly from the original record when converted from time to depth. However, sound velocities may vary with lateral position as well as with depth, and all variations must be known to construct a proper sub-surface model.

Sound Velocity in Strait of Georgia Water

Under normal oceanic conditions, the speed of sound in sea water varies between the limits of 1,400 to 1,560 meters per second (4,600 to 5,100 feet per second). For accurate measurements, corrections must therefore be made to account for these variations. Since variations are a function of the in situ salinity, temperature and pressure distribution, these parameters provide the necessary means for an accurate correction to the sea water depths. Officer (1958, p. 19) gives the following approximate correction factors for

water differing from a standard sea water: an increase of 3.13 meters per second per Centigrade degree (5.7 feet per second per Fahrenheit degree); an increase of 1.34 meters per second (4.27 feet per second) per one thousandth part increase in salinity; an increase of 1.82 meters per second per 100 meter depth increment.

In the Strait of Georgia, a high influx of fresh water from precipitation and river runoff leads to the formation of a brackish surface layer over much of the Strait (Waldichuk, 1952). During spring freshet, surface salinity varies laterally from two parts per thousand near the Fraser River mouth to twenty-six parts per thousand in the southern Strait. During the months of December to April when river discharge is at a winter minimum, surface salinity varies from twenty to thirty parts per thousand over the same area. Below 30 meters (100 feet) in depth, the seasonal variation is negligibly small.

Temperature variations follow a similar pattern. In summer the upper brackish layer is warm and a strong, shallow thermocline exists. Winter temperatures are more uniform in both lateral and vertical directions (Waldichuk, 1953).

The effect of these variations on the speed of sound can be assessed for the survey area. Assuming a two layer model from the salinity and temperature data of Waldichuk can approximate the winter conditions in the Strait, since most

of the survey was carried out in January, the following data are obtained. The surface layer is 15 meters (50 feet) thick with an average temperature of 6.1°C and salinity at 26 parts per thousand. The bottom layer is 366 meters (1,200 feet) thick with temperature of 7.2°C and salinity at 30 parts per thousand. Lateral and vertical variations at that time are assumed to have been near the yearly minimum, thus a single two layer model will typify the whole area. Using tables of the U.S. Naval Hydrographic Office, Publication Number 614, the respective layer velocities are 1,460 meters (4,785 feet) per second and 1,475 meters (4,840 feet) per second. As the sound wave traverses a much thicker lower layer, it is apparent that true water depth is deeper than the records indicate since they are calibrated for a sound speed of 1,465 meters (4,800 feet) per second. The difference, however, is much less than one percent and is therefore negligible.

Sound Velocities in Unconsolidated Sediments

Data for sound velocities in sediments of the study area is lacking for most of the units encountered. Deep wells drilled by oil companies have been logged only in their lower depth intervals while those intervals within the range of the continuous seismic profiling equipment are not logged.

An attempt was made to measure sound velocity in unconsolidated beach sands both at Point Grey and at Boundary

Bay. Hammer seismic equipment was used at low tide when the sand had not yet dried out. Results varied over individual hammer blows but gave a mean velocity in wet, coarse sand of 1,525 meters (5,000 feet) per second. For comparison, a test in loose, dry sand farther up the beach gave a velocity of 457 meters (1,500 feet) per second. In wet, fine sand and silt at Boundary Bay the average velocity was 1,370 to 1,400 meters (4,500 to 4,600 feet) per second although individual attempts ranged over much wider limits.

Clark's "Handbook of Physical Constants" (1966) quotes a velocity range of 500 to 2,000 meters (1,640 to 6,600 feet) per second in alluvium at shallow depth. For a depth in alluvium of 2,000 meters (6,600 feet), the velocity range quoted is 3,000 to 3,500 meters (9,800 to 11,500 feet) per second.

A refraction shot in alluvium of the Salmon River delta on northern Vancouver Island (White, 1962) gave a velocity measurement of 1.58 kilometers (5,200 feet) per second. The alluvium is described (p.65) as "river deposited silt".

The private consulting firm of Swan Wooster-CBA (a joint venture) made available (personal communication, 1969) results of seismic velocity measurements in three boreholes in Vancouver Harbour just east of First Narrows. The boreholes penetrated through coarse deltaic and glacial materials

and of the three, two bottomed in Tertiary sandstone. The velocity was found to range from 1,110 to 1,830 meters (3,650 to 6,000 feet) per second for sound in deltaic sediments. The material is composed of fine, medium and coarse sands, sand and gravel, silty sands, silts, shells, and organic material, described as ranging from compact to dense. The higher velocities appear in the denser material.

From the above information, the velocity of sound in the upper Fraser Delta is most likely to be near 1,525 meters (5,000 feet) per second. Since this velocity is close to that of water, sediment thickness read directly off the profiles at the water velocity will err on the low side by approximately 4%.

No velocity measurements are available for the lower delta front sediments or those of the deep basins. Oil companies which have worked in similar sedimentary frameworks assume a velocity of 1,830 meters (6,000 feet) per second (P. Fuller, Shell Canada Ltd., personal communication, 1969). As this velocity is apparently not based upon measurement, its validity is questionable. Since recent hemipelagic sedimentation is involved at these sites, data from deep sea and continental shelf measurements may be applicable. Laboratory measurements made by Hamilton (1965) on deep sea clayey silts and silty clays, when corrected to in situ values, ranged from 1,500 to 1,555 meters (4,900 to 5,100

to 1,555 meters (4,900 to 5,100 feet) per second for depths of burial to 140 meters (460 feet). In situ measurements on continental shelf sediments (Hamilton, et al, 1956) gave velocities of 1,750 meters (5,750 feet) per second for coarse sands decreasing to 1,680 meters (5,520 feet) per second for fine sands, and dropping to 1,460 meters (4,800 feet) per second for finer sandy-silt and clayey-silt material. Thus average velocities for the lower delta front may be near 1,585 meters (5,200 feet) per second while for the clayey muds of the basins, a velocity closer to that of water is not unreasonable. In fact, the basin sediments may be, in the surface layer at least, high porosity sediments, in which case sound velocity may be less than that of water (Hamilton, 1956). However, as velocity increases with depth of burial, and basin sediments reach several hundred meters thickness, an average velocity of 1,585 meters (5,200 feet) per second for these sediments as well has been assumed. Thus the thickness as read from the profiles has been increased by approximately eight percent. The hemi-pelagic sediments on the ridge tops and sides may have similar sound velocities.

Velocities in Pleistocene drift and till and intertill sediments are not known. The Vancouver Harbour boreholes which passed through glacial materials give a range of velocities of 1,220 to 2,740 meters (4,000 to 9,000 feet) per second. A similar range was obtained in Pleistocene sediments

from two boreholes logged in the Fraser Valley (Richfield Pure Abbotsford; Richfield Pure Sunnyside). In the study area, the variation of character of the seismic record over Pleistocene deposits suggests similar variations in the sediments. Thus no single velocity is likely to be suitable for all cases. However, a velocity one-third greater than the velocity of sound in sea water, 1,950 meters (6,400 feet) per second, is probably a realistic average about which the true velocity will vary. This value has been used for measurements of Pleistocene sediment thickness in this thesis.

Sound Velocities in Bedrock

Bedrock velocities have been determined for some rock units near the study area (White and Savage, 1965), but since little bedrock penetration was achieved in this survey, only velocities in the Whatcom Basin sediments and Upper Cretaceous Nanaimo Group formations are required.

The two boreholes in Vancouver Harbour which terminated in Tertiary sandstones indicate velocities of 20% and 2,660 meters (6,800 and 8,750 feet) per second in these rocks. Two other wildcat wells reaching Tertiary sandstones in Whatcom Basin for which velocity logs are available (Richfield Pure Abbotsford; Richfield Pure Sunnyside) give results ranging from 3,360 to 4,730 meters (11,000 to 15,500 feet) per second. Since the latter two logs were recorded deep within the wells, the depth of burial may be a factor in the increased velocities

obtained. For this thesis, a velocity of 2,560 meters (8,400 feet) per second was adopted for Tertiary bedrock.

Refraction shots in the Strait of Georgia have provided an average compressional wave velocity of 4.05 kilometers (13,300 feet) per second for Upper Cretaceous Nanaimo Group rocks (Milne and White, 1960). Wells that penetrated Upper Cretaceous rocks (Richfield Pure Abbotsford; Richfield Pure Sunnyside; Richfield Pure Point Roberts) give sonic velocities ranging from 4.05 to 6.1 kilometers (13,300 to 20,000 feet) per second. Because the value of 4.05 kilometers per second was obtained on the western side of the Strait as a working value, it is used for sound velocity in Upper Cretaceous rocks in this study.

LEGEND

HOLOCENE SEDIMENTS

H_i UNDIFFERENTIATED

H_t TRANSPARENT TYPE

H_3 MODERN DELTA

H_2 MODERN PRO-DELTA

H_1 ANCIENT PRO-DELTA

PLEISTOCENE SEDIMENTS

P_{mr} McCALL RIDGE UNIT

P_{rs} ROBERTS SWELL UNIT

P_s STRATIFIED UNDIFFERENTIATED

P_n NON-STRATIFIED OR CHAOTIC

P UNDIFFERENTIATED

Q PLEISTOCENE AND/OR HOLOCENE SEDIMENTS

BEDROCK

B_t INFERRED TERTIARY, MAY INCLUDE SOME UPPER CRETACEOUS

B_k INFERRED UPPER CRETACEOUS, MAY INCLUDE SOME TERTIARY

B_j PROBABLE COAST RANGE INTRUSIVE, VOLCANIC OR PRE-UPPER CRETACEOUS METASEDIMENTARY ROCK

B UNDIFFERENTIATED BEDROCK

LEGEND

HOLOCENE SEDIMENTS

H_1 UNDIFFERENTIATED

H_t TRANSPARENT TYPE

H_3 MODERN DELTA

H_2 MODERN PRO-DELTA

H_1 ANCIENT PRO-DELTA

PLEISTOCENE SEDIMENTS

P_{mr} McCALL RIDGE UNIT

P_{rs} ROBERTS SWELL UNIT

P_s STRATIFIED UNDIFFERENTIATED

P_n NON-STRATIFIED OR CHAOTIC

P UNDIFFERENTIATED

Q PLEISTOCENE AND/OR HOLOCENE SEDIMENTS

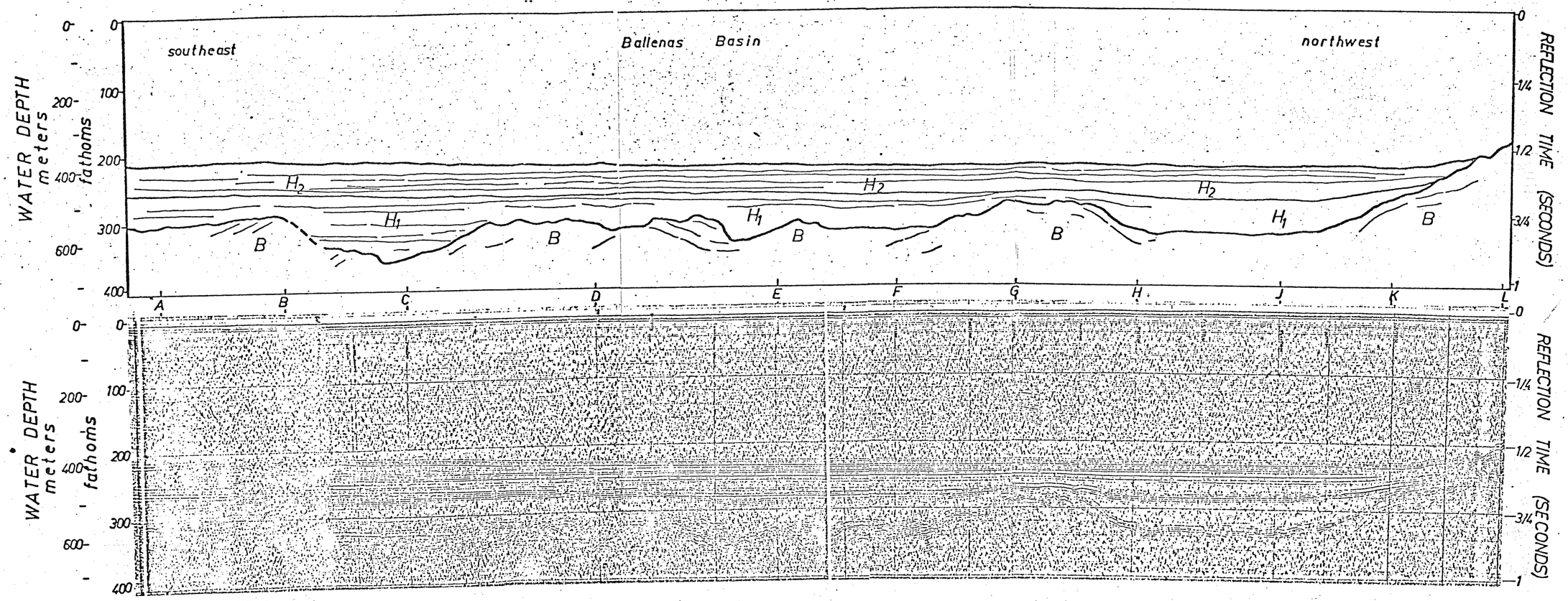
BEDROCK

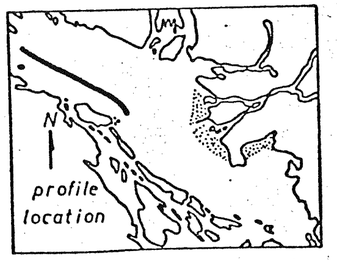
B_t INFERRED TERTIARY, MAY INCLUDE SOME UPPER CRETACEOUS

B_k INFERRED UPPER CRETACEOUS, MAY INCLUDE SOME TERTIARY

B_j PROBABLE COAST RANGE INTRUSIVE, VOLCANIC OR PRE-UPPER CRETACEOUS METASEDIMENTARY ROCK

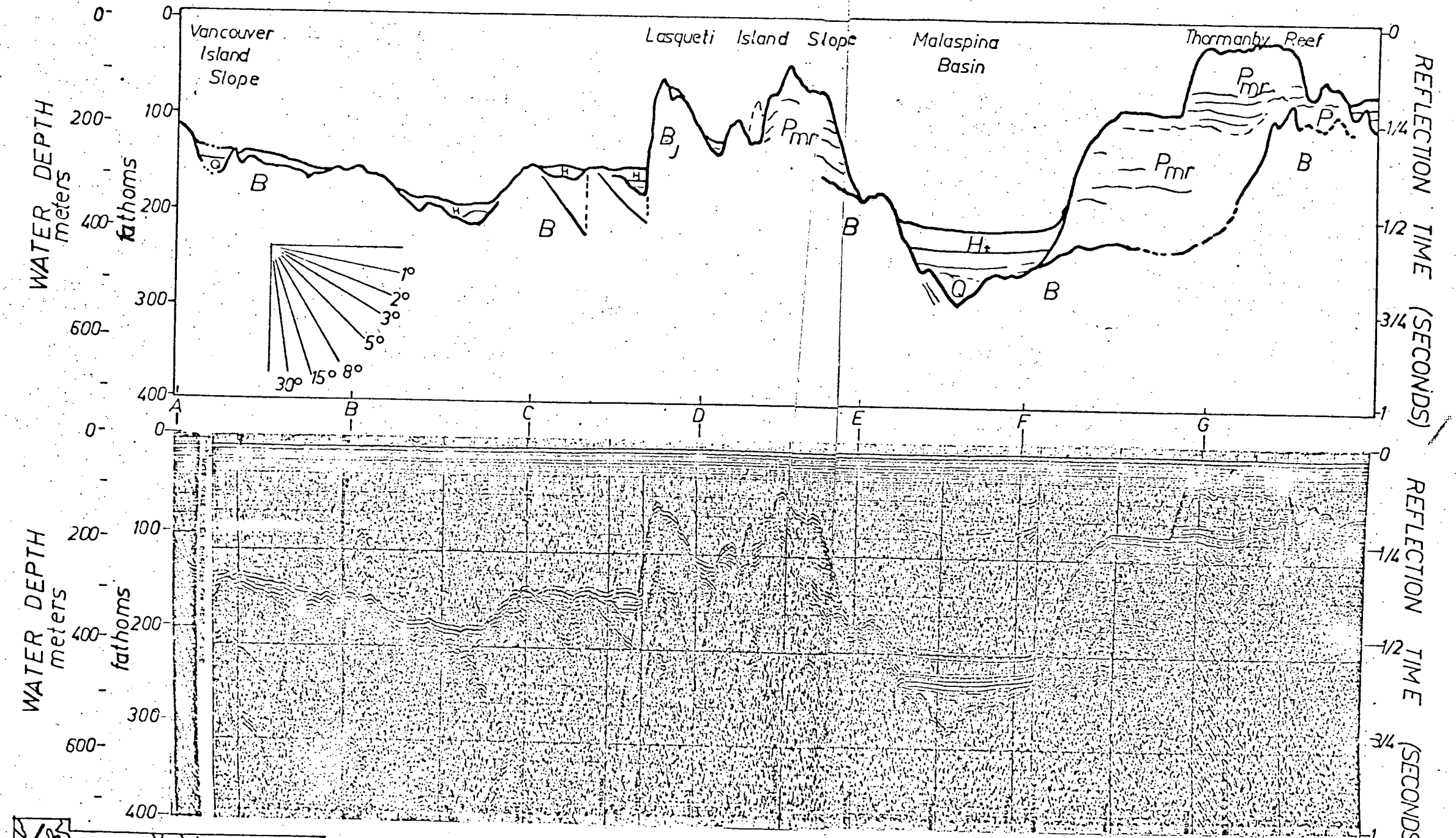
B UNDIFFERENTIATED BEDROCK

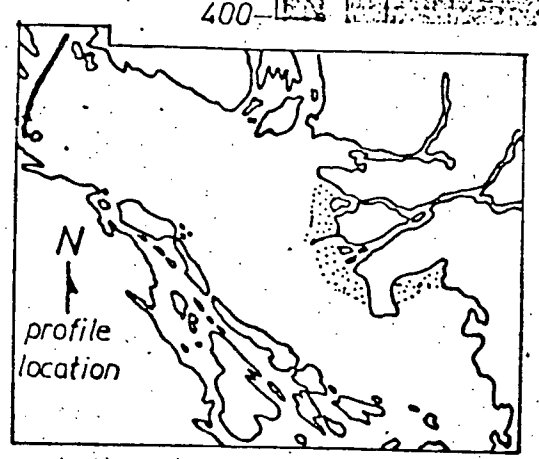




LENGTH 46 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

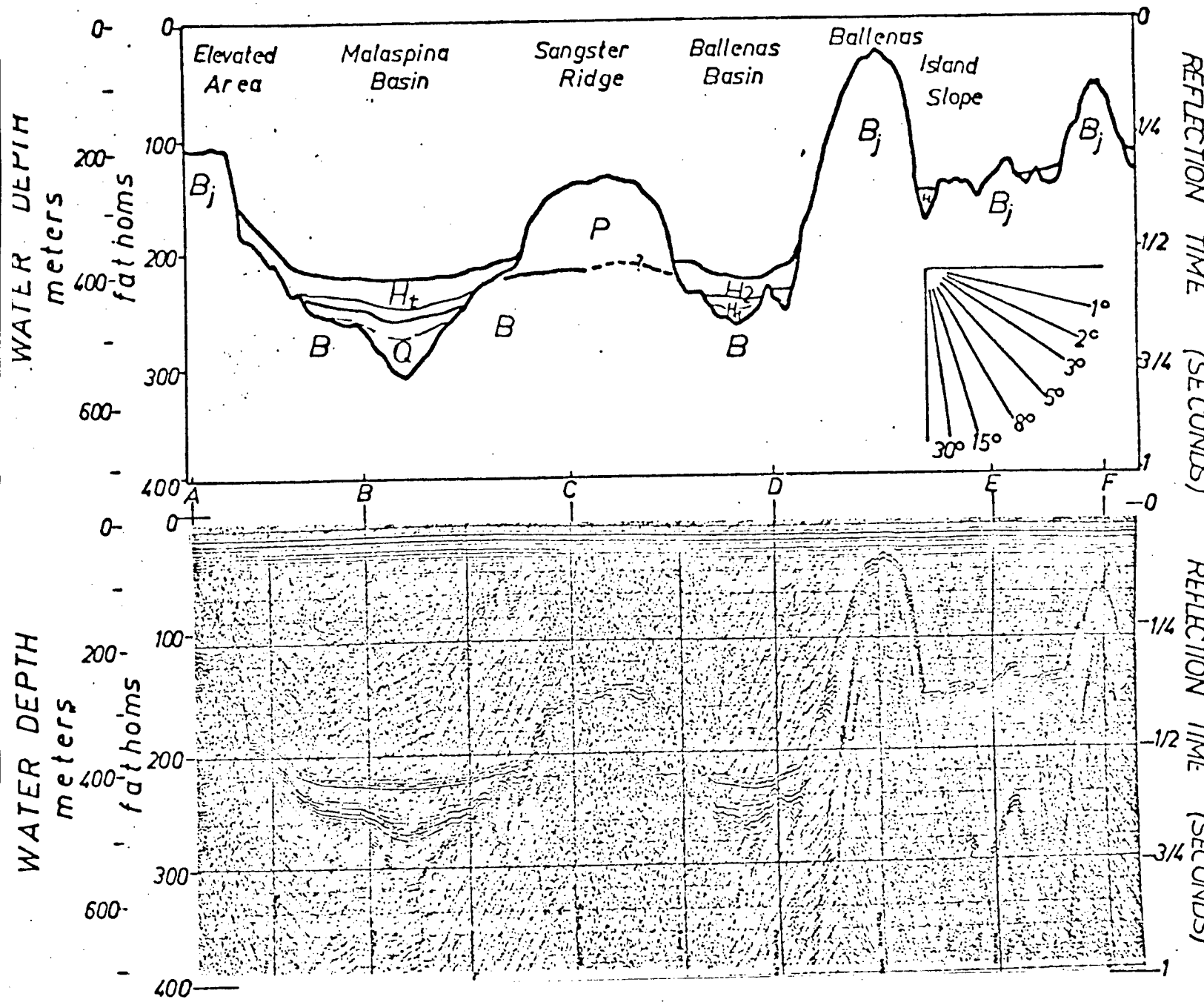
PLATE I

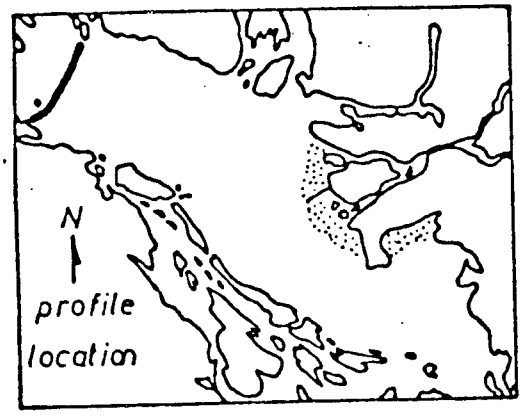




LENGTH 25 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

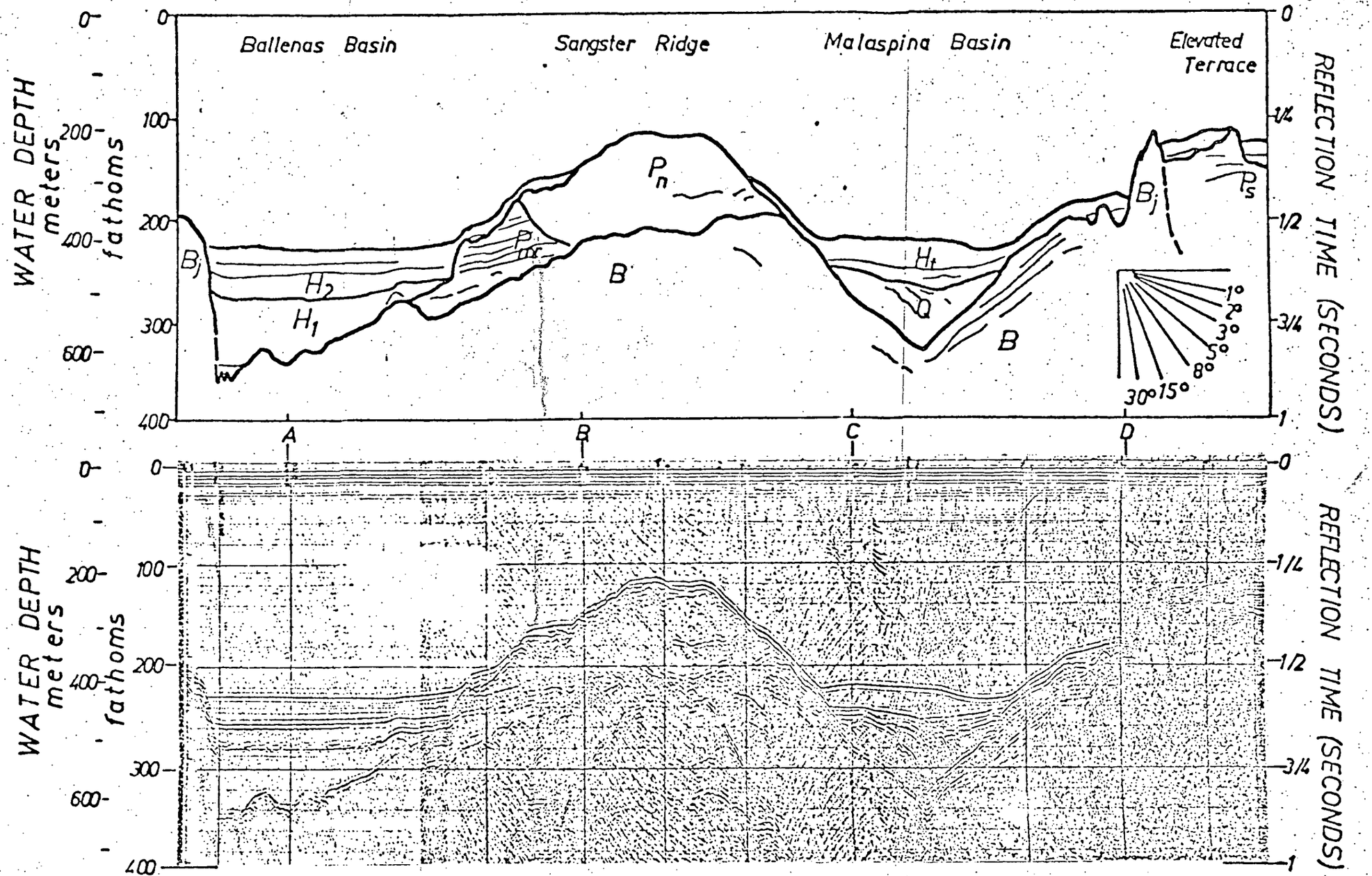
PLATE II

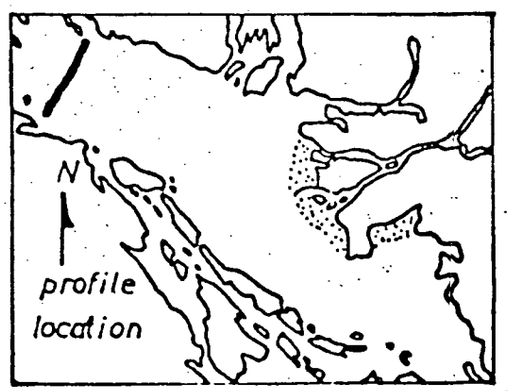




LENGTH 22 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

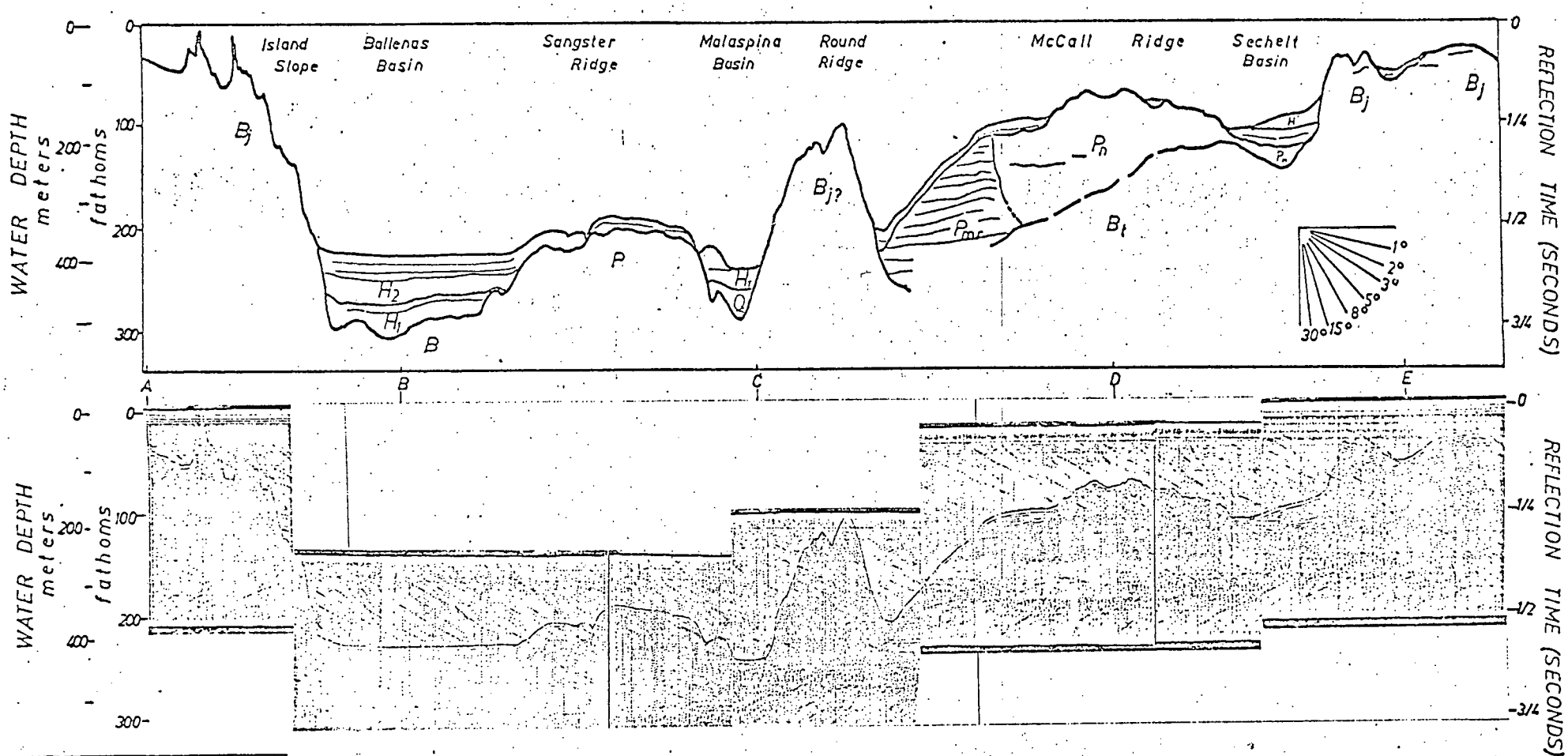
PLATE III

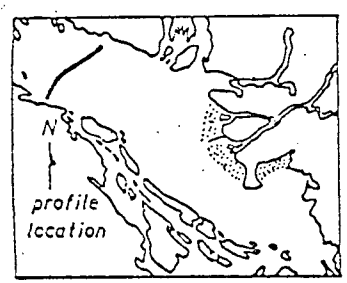




LENGTH 16 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

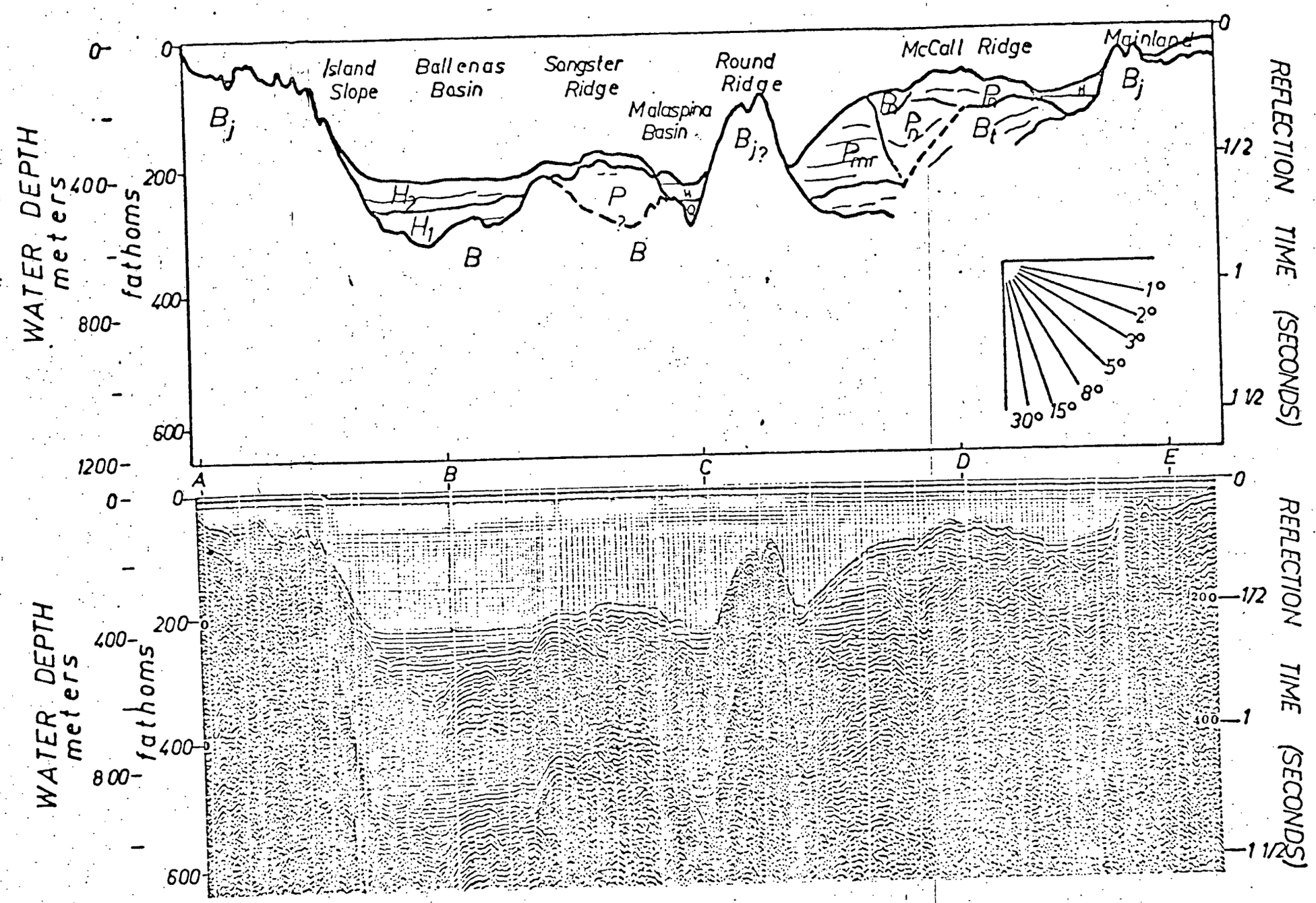
PLATE IV

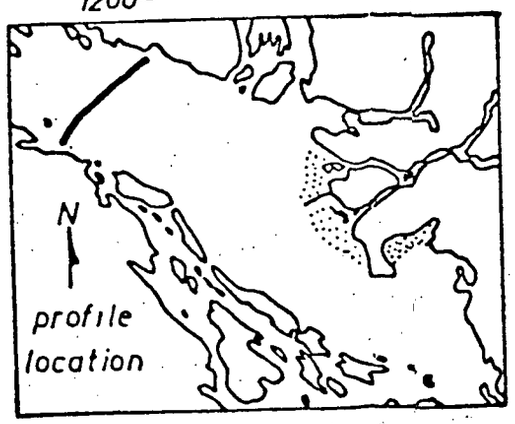




SPARKER PROFILE BY
SHELL CANADA LIMITED
LENGTH 31 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

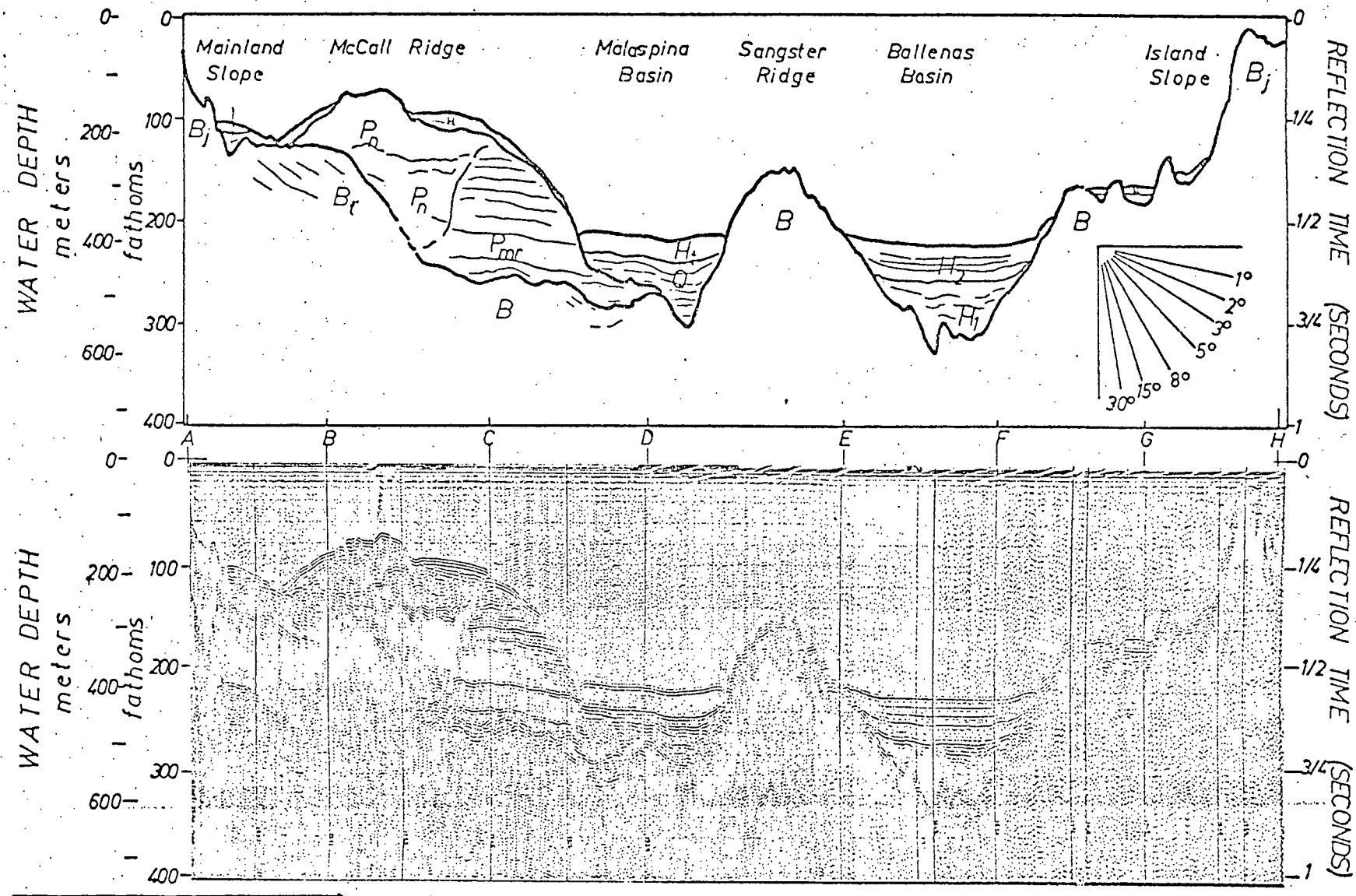
PLATE V

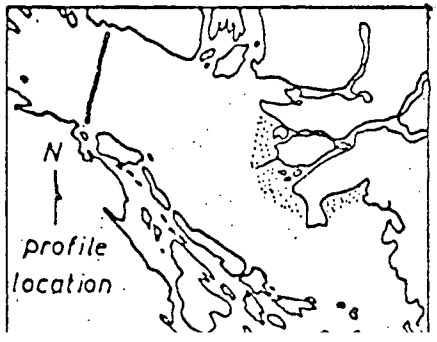




GAS EXPLODER PROFILE BY SHELL CANADA LIMITED.
LENGTH 31 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

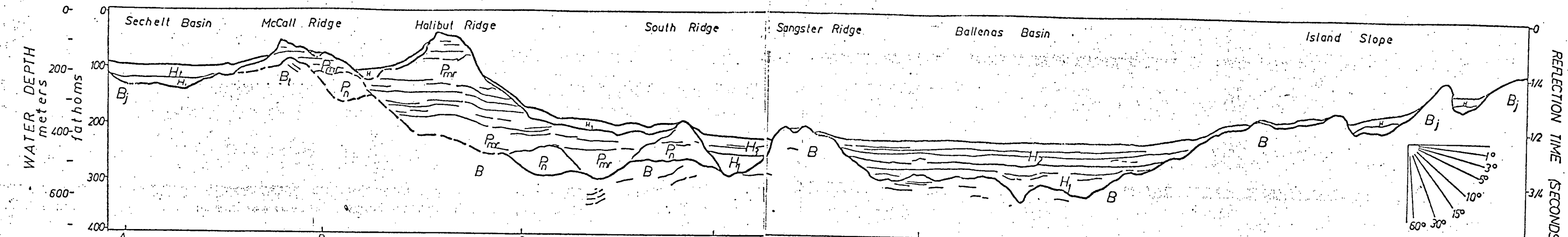
PLATE VI

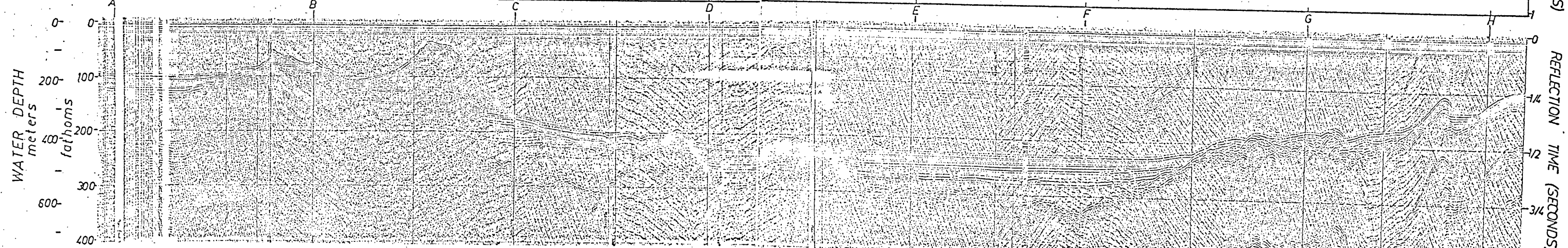


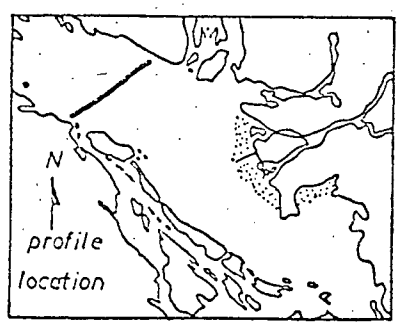


LENGTH 27 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

PLATE VII

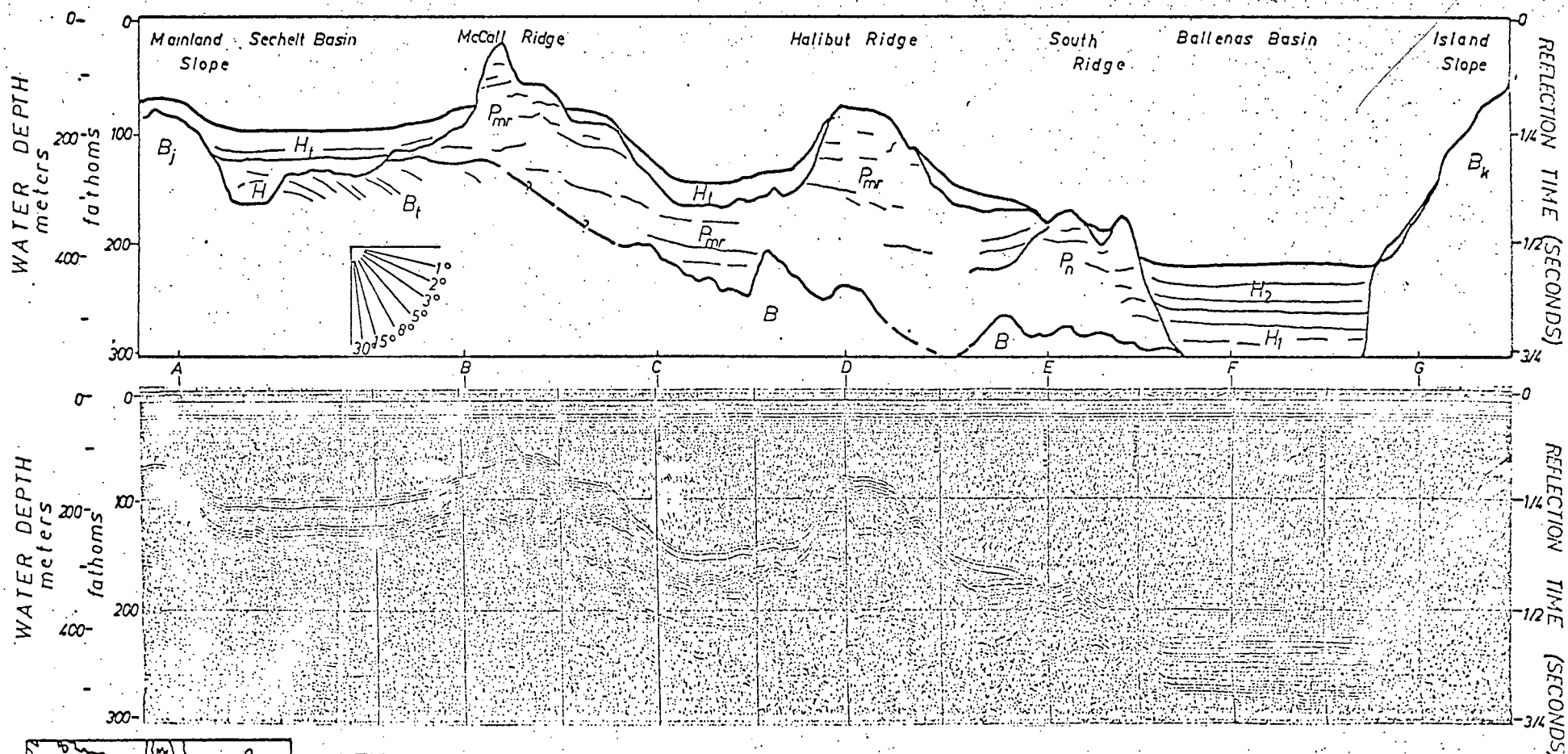


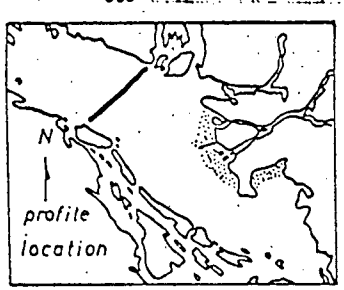




LENGTH 28 KILOMETERS
POSITIONS ON FIGURE 5.
LEGEND OPPOSITE PLATE I

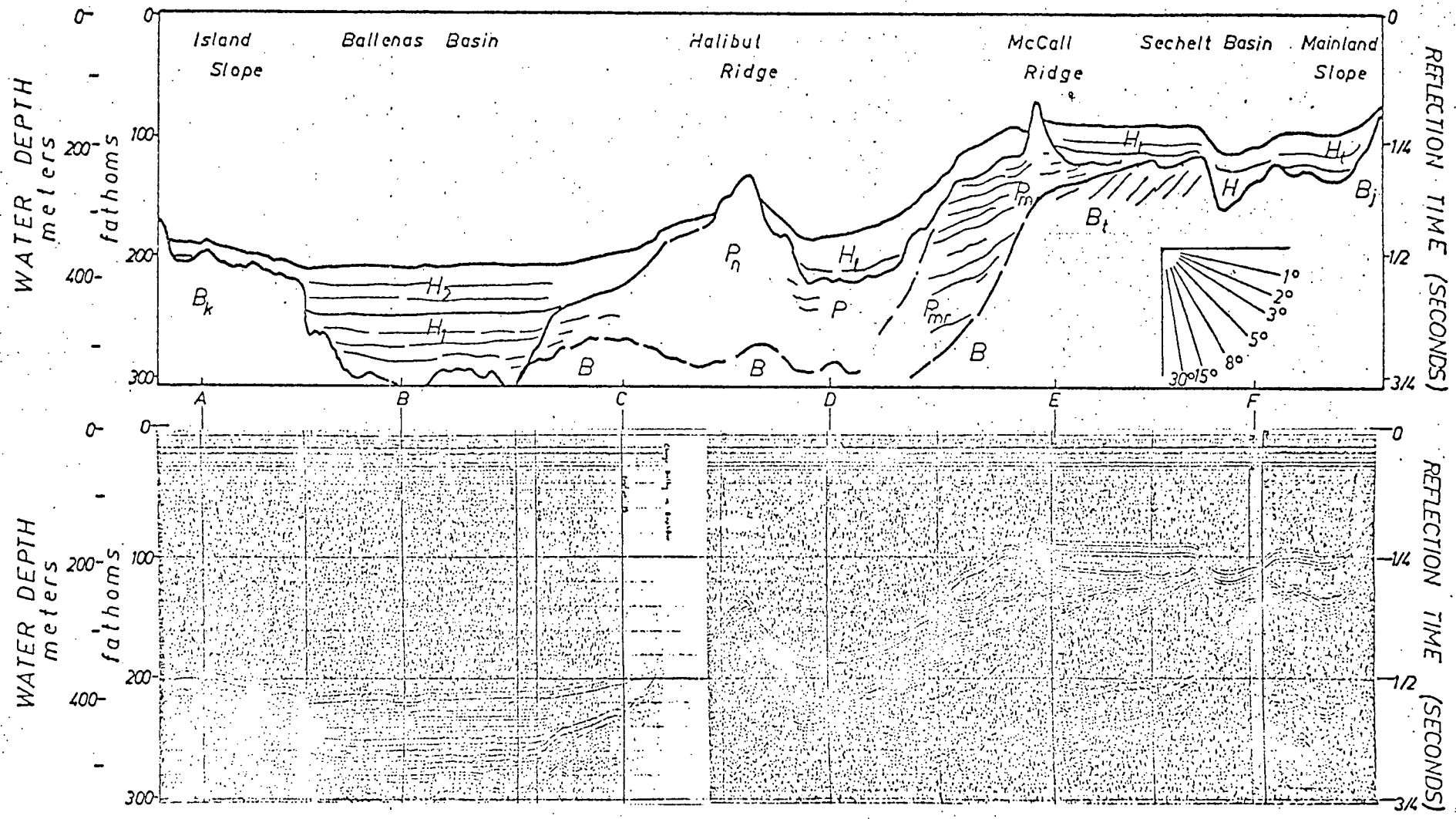
PLATE VIII

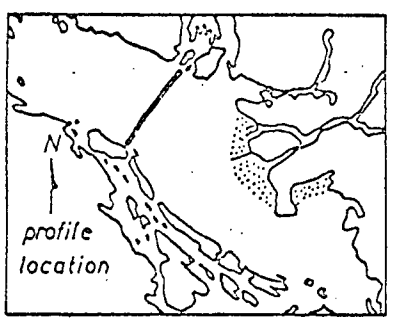




LENGTH 27 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

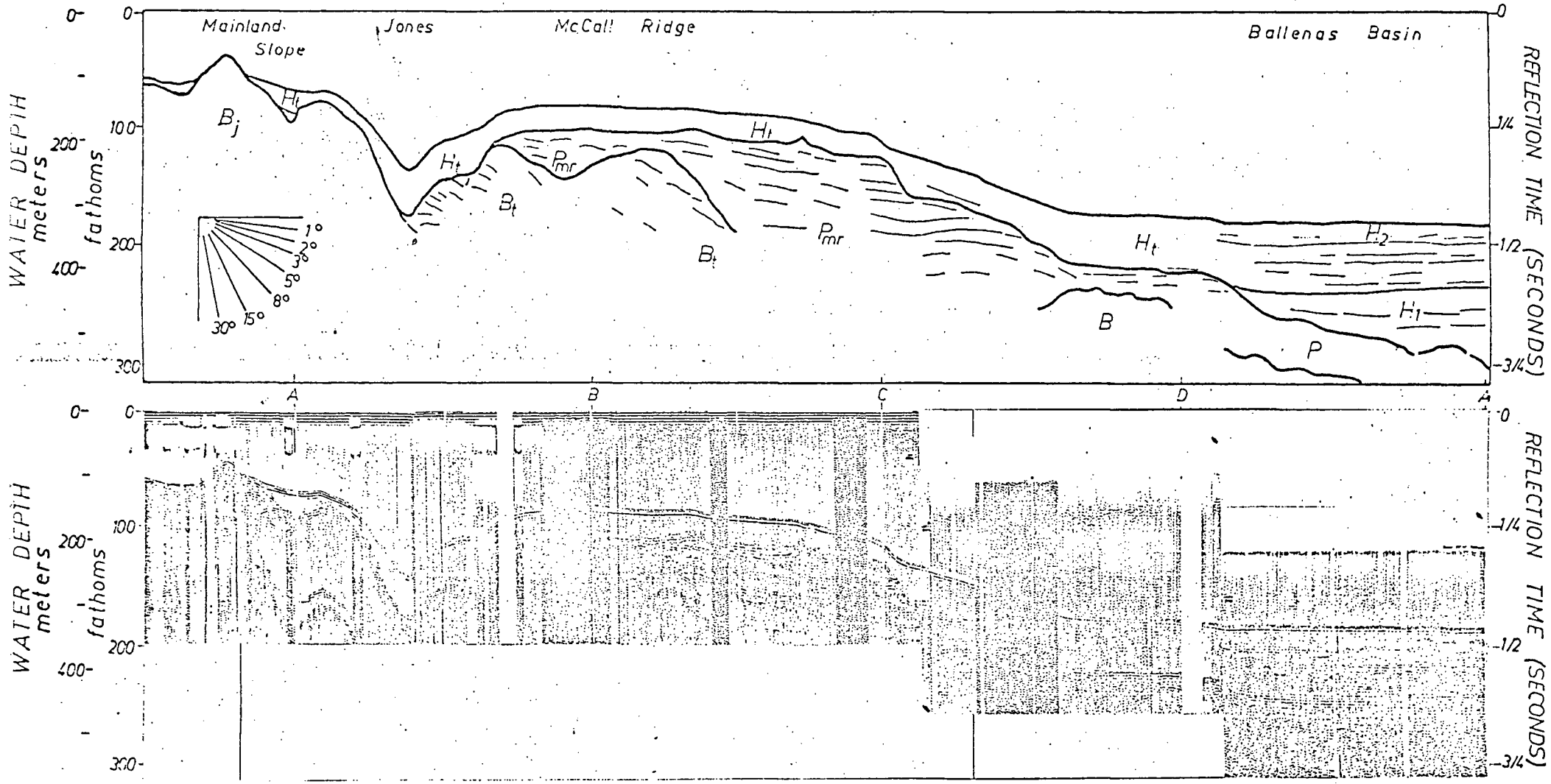
PLATE IX

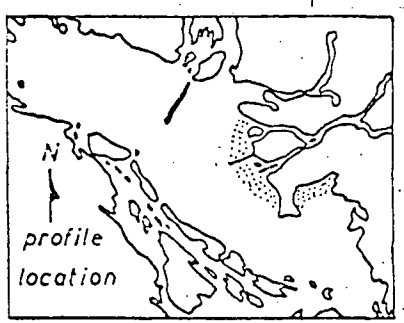




LENGTH 24 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

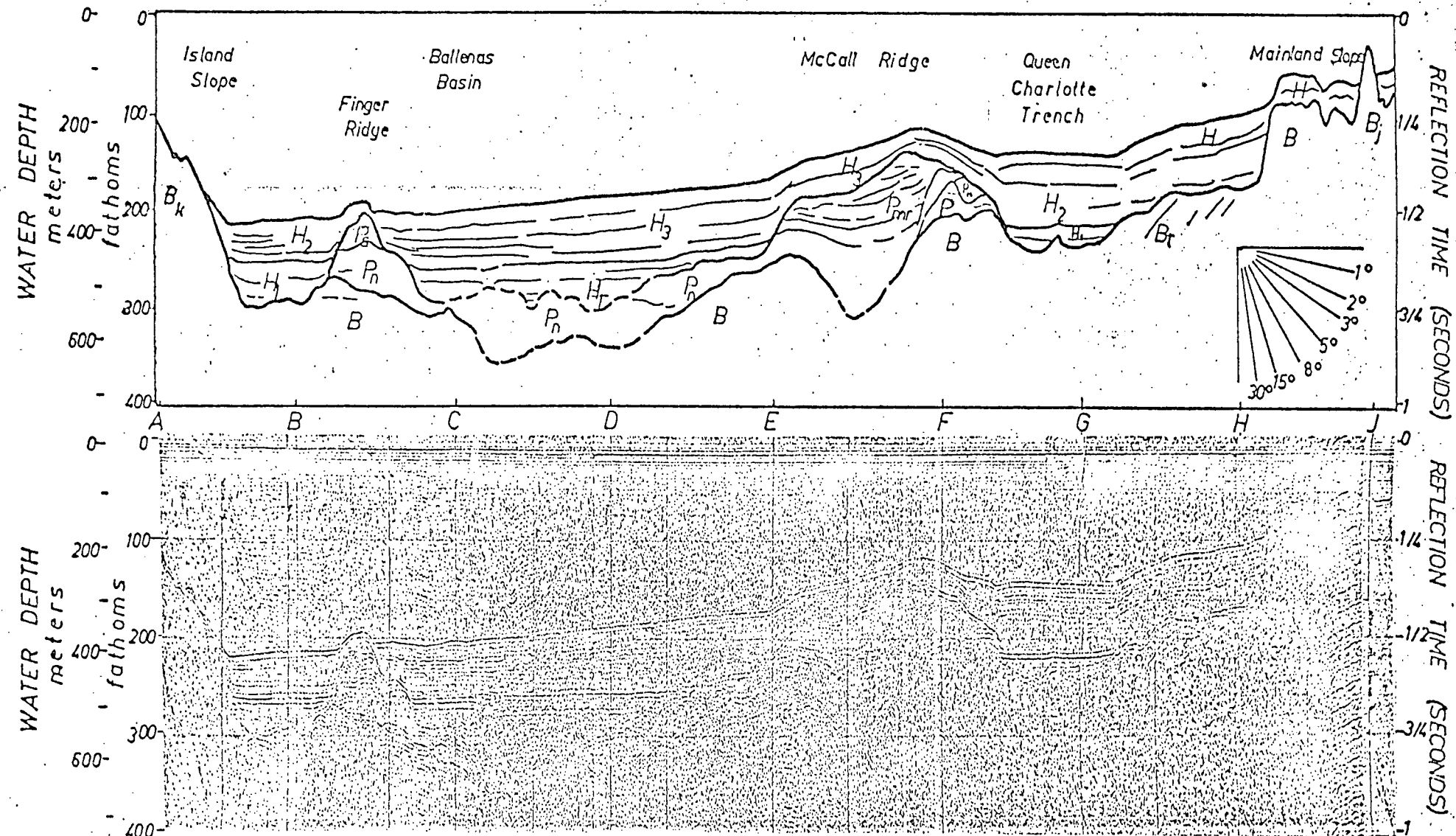
PLATE X

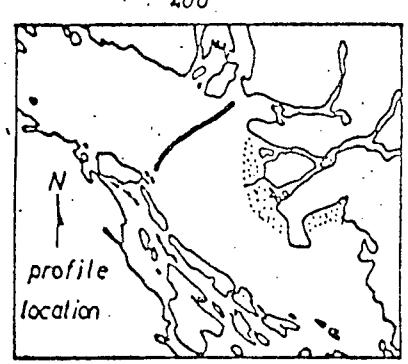




LENGTH 14 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I
SPARKER PROFILE BY
SHELL CANADA LIMITED

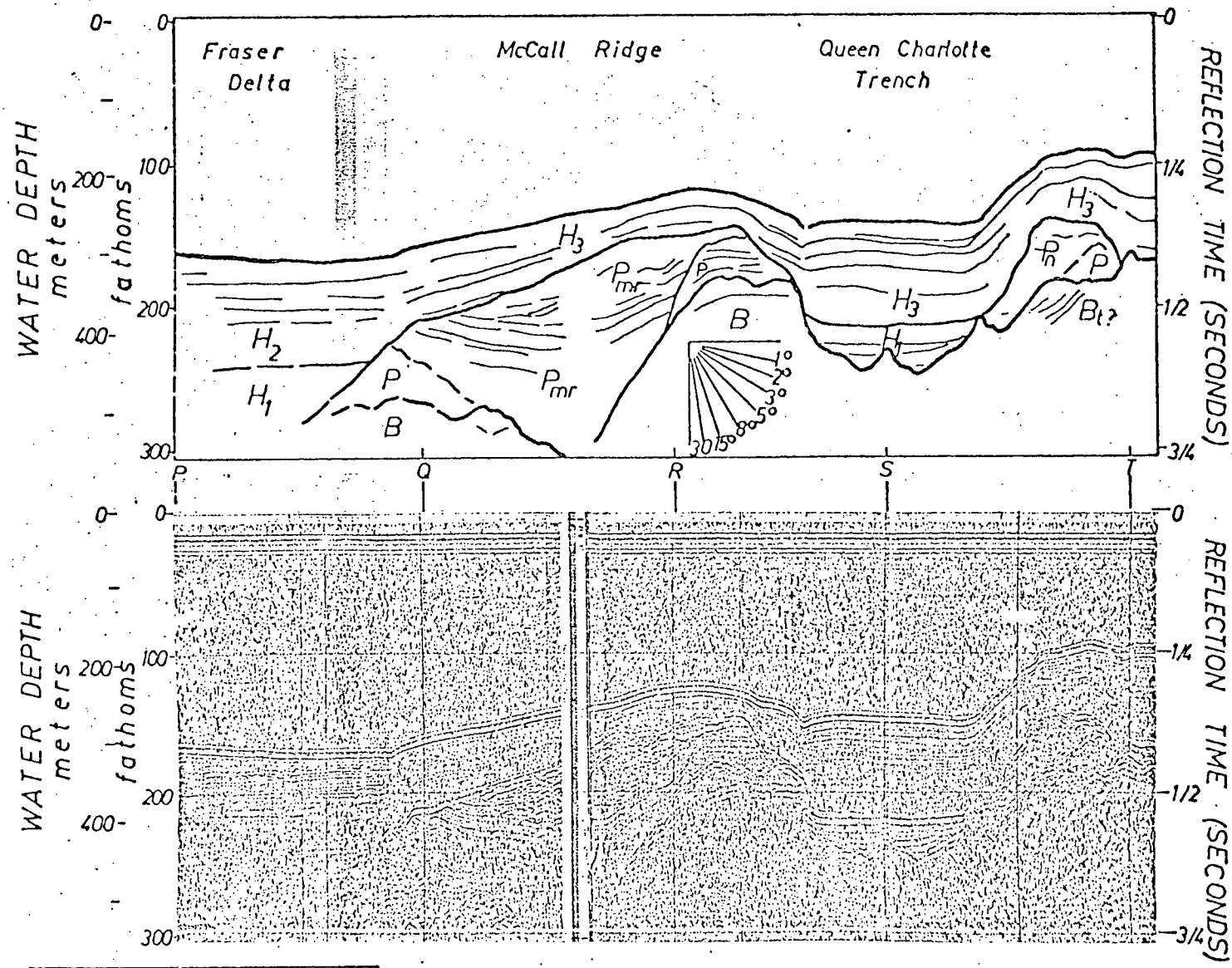
PLATE XI

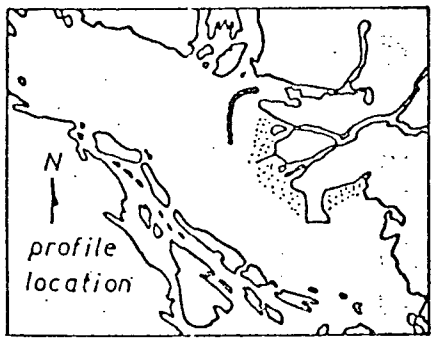




LENGTH 29 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

PLATE XII



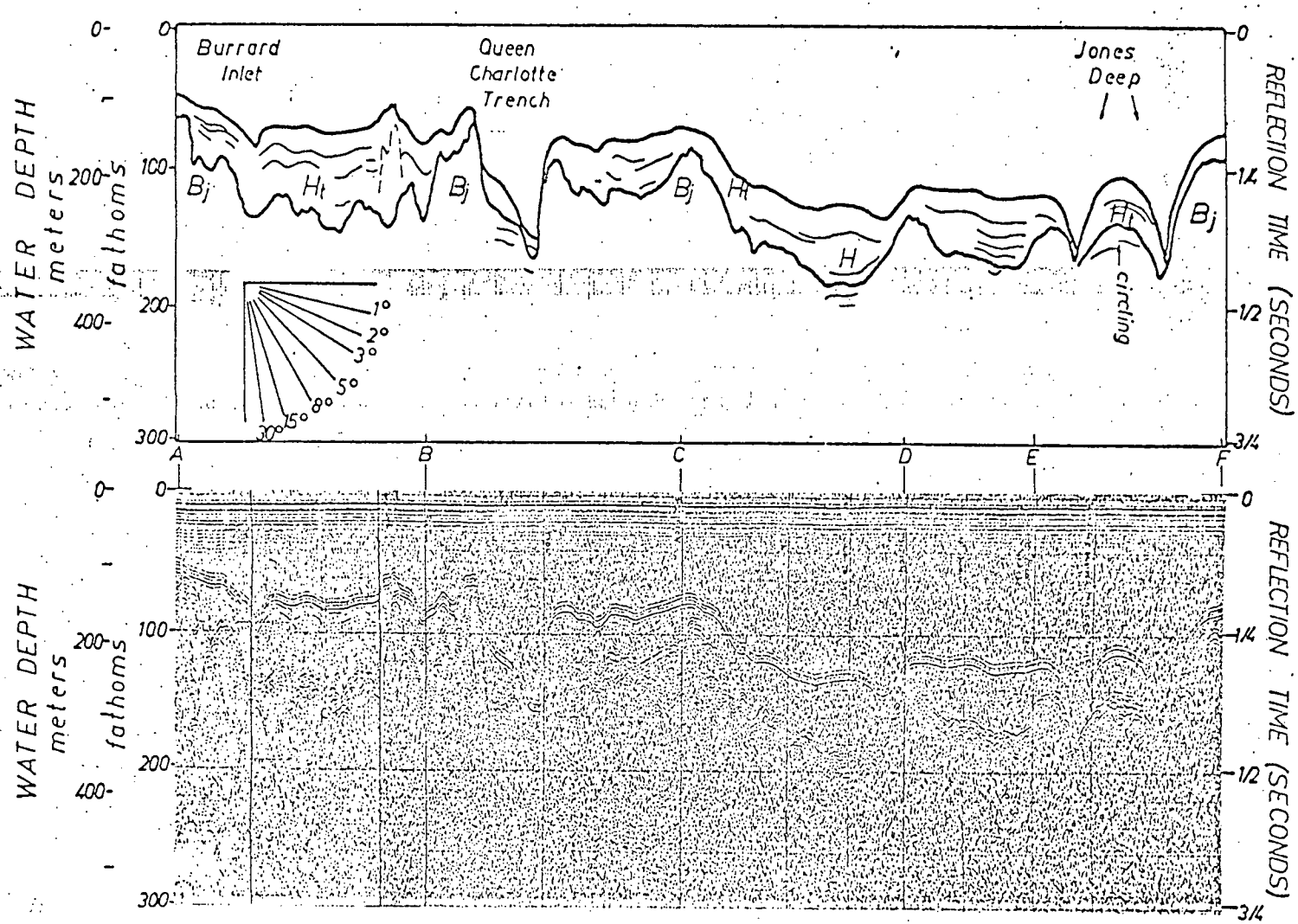


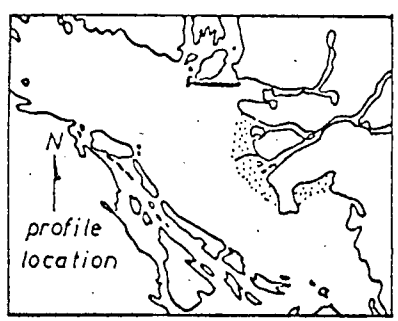
LENGTH 15 KILOMETERS

POSITIONS ON FIGURE 5

LEGEND OPPOSITE PLATE I

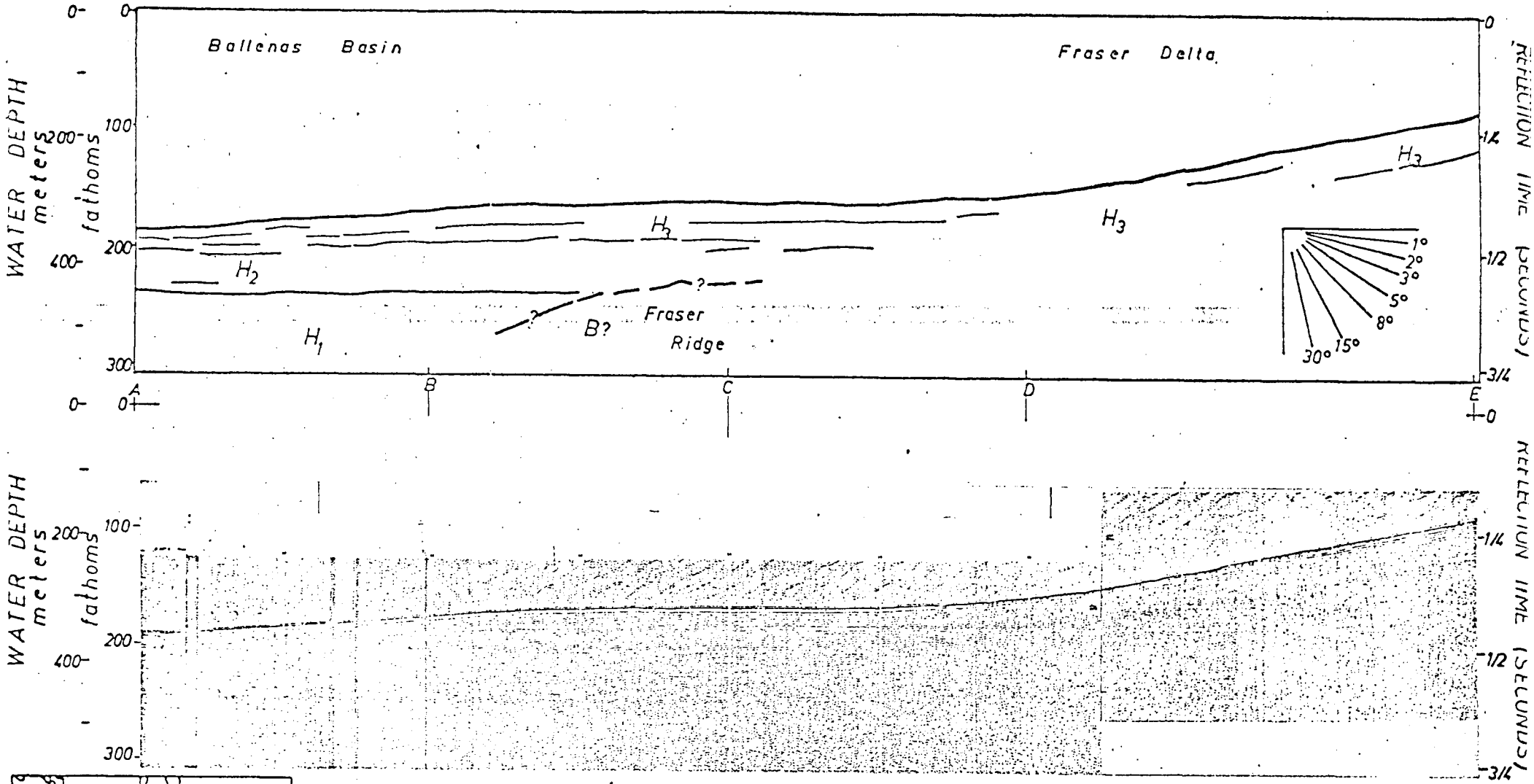
PLATE XIII

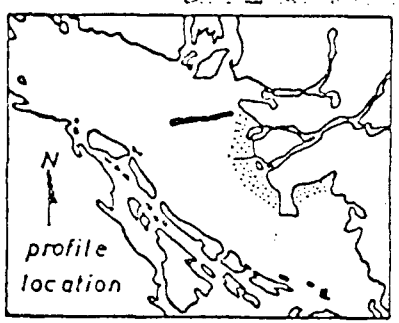




LENGTH 16 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

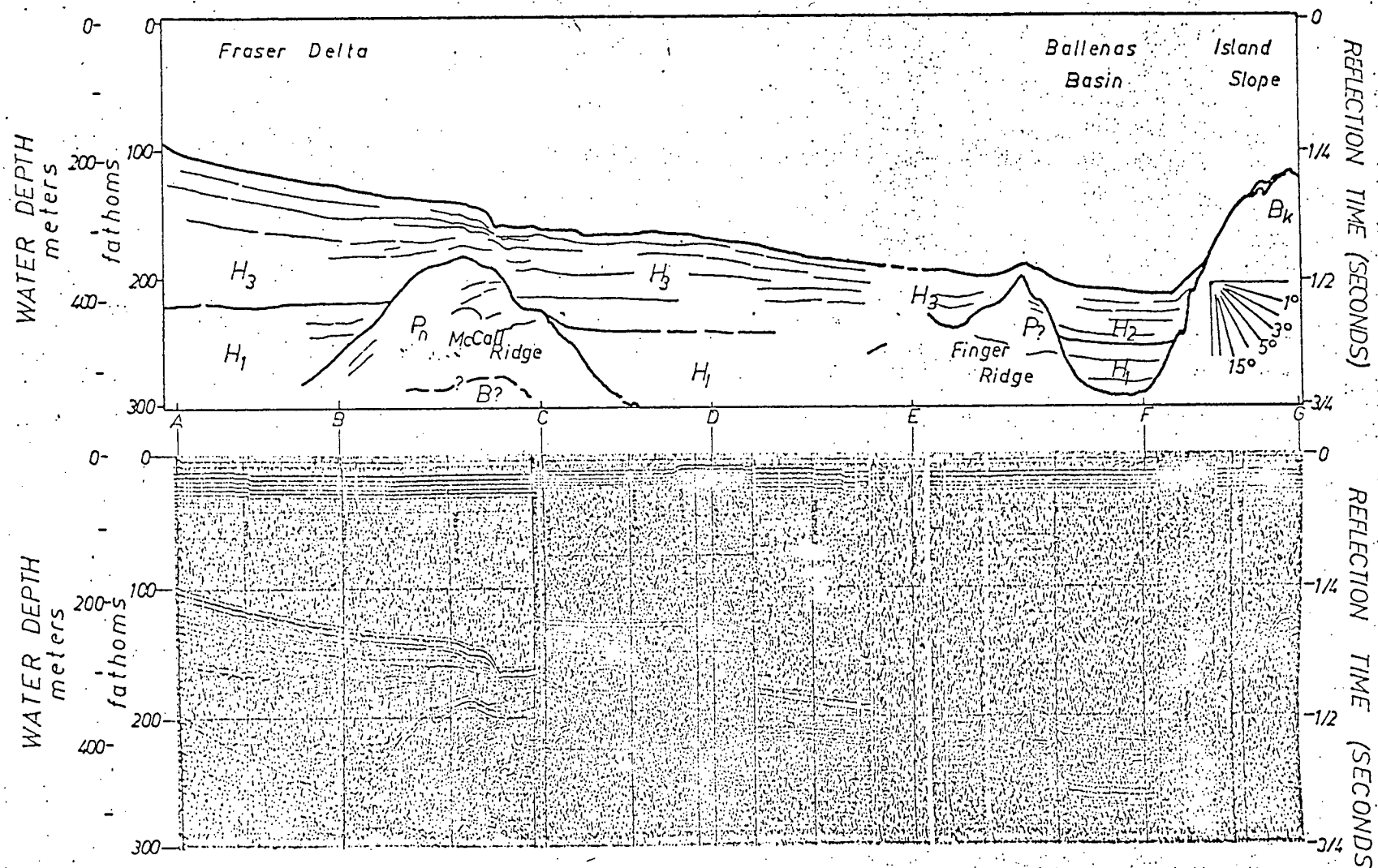
PLATE XIV

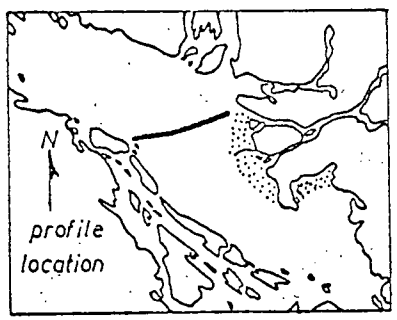




LENGTH 15 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I
SPARKER PROFILE BY
SHELL CANADA LIMITED

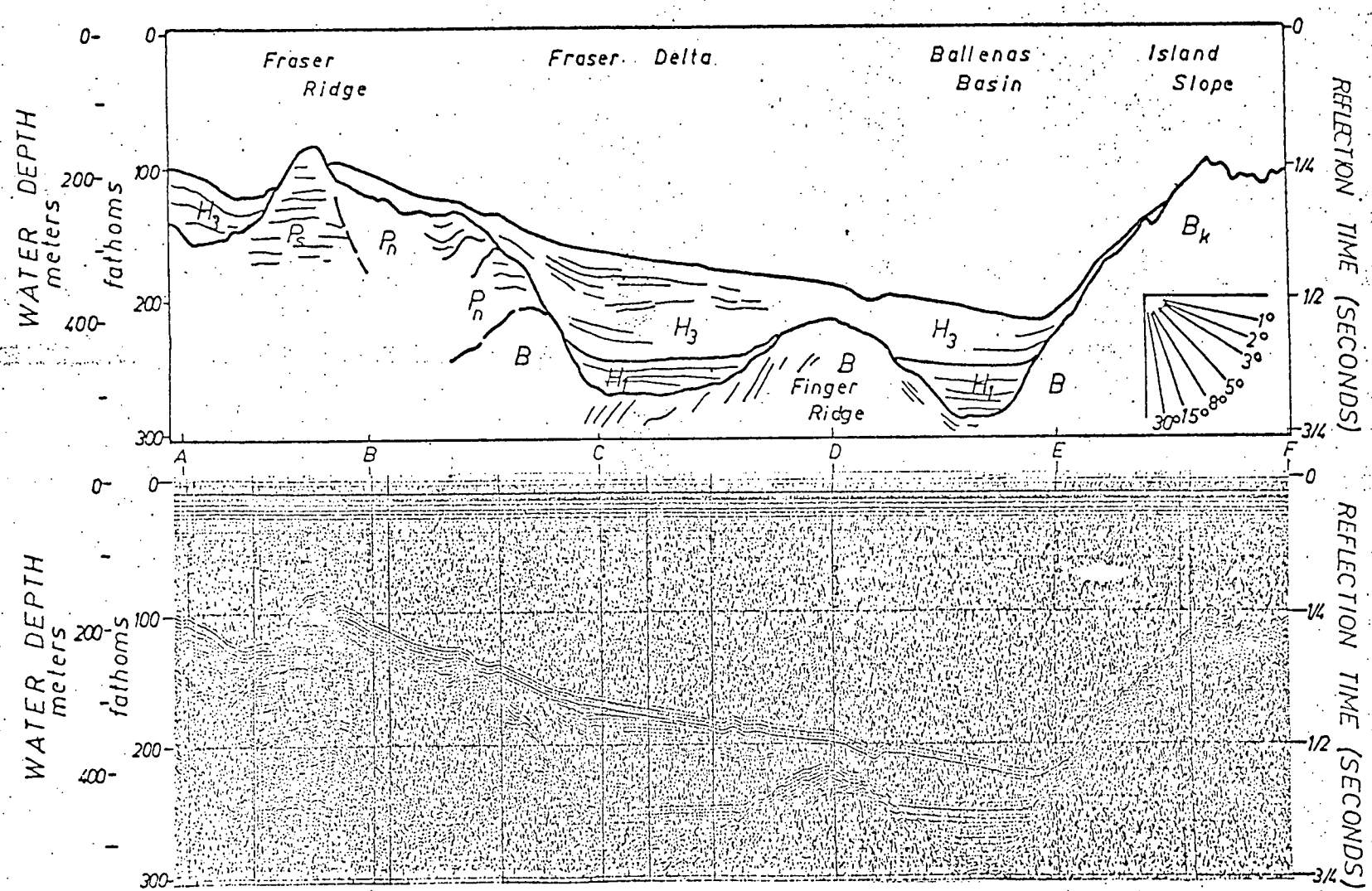
PLATE XV

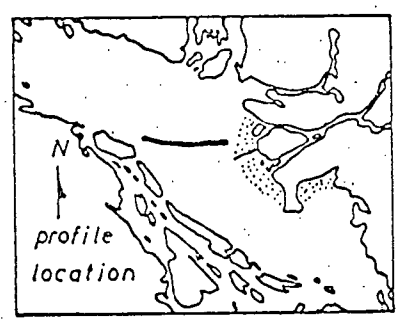




LENGTH 24 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

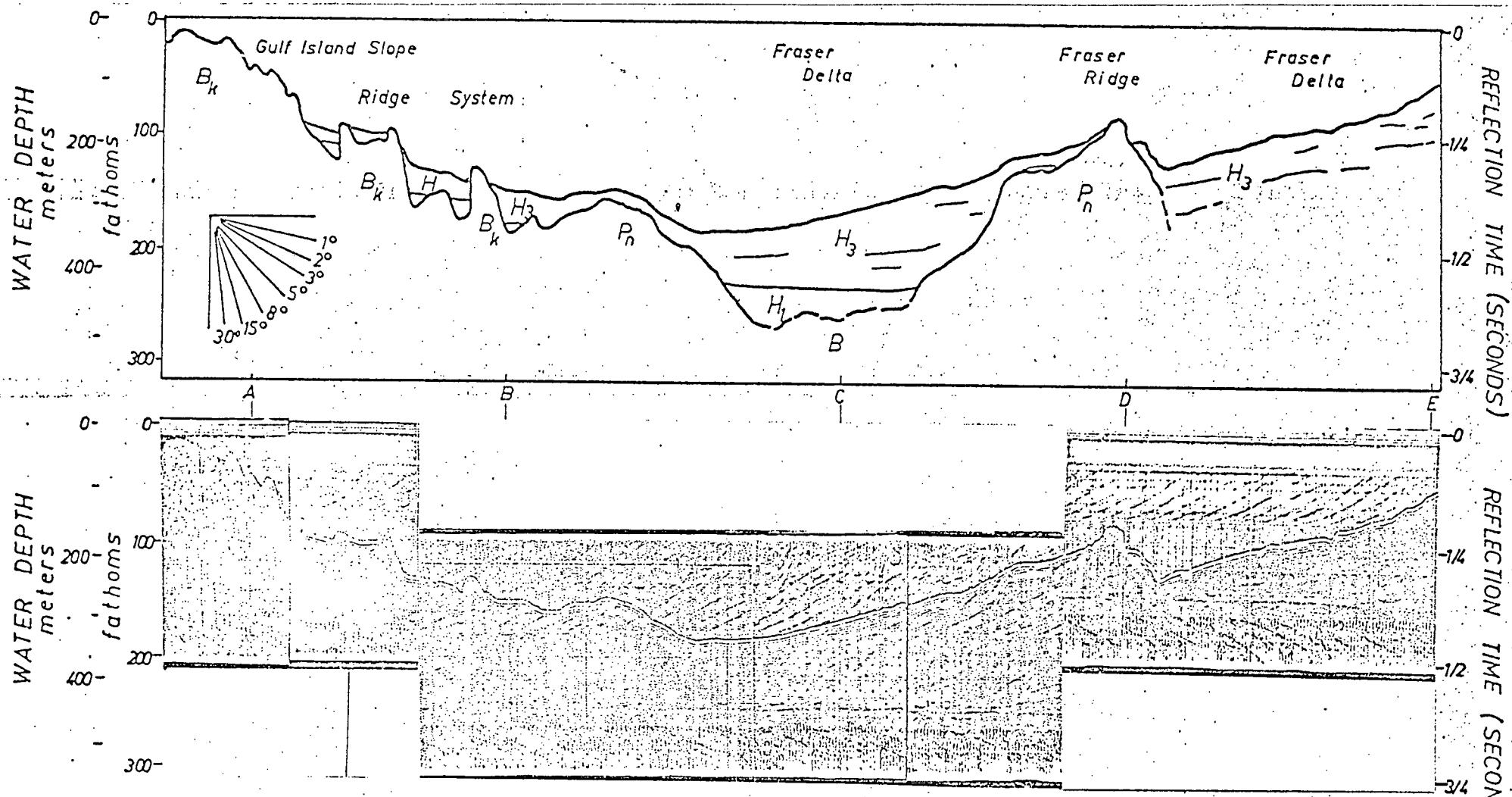
PLATE XVI

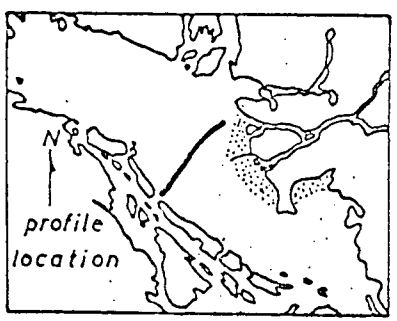




LENGTH 34 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

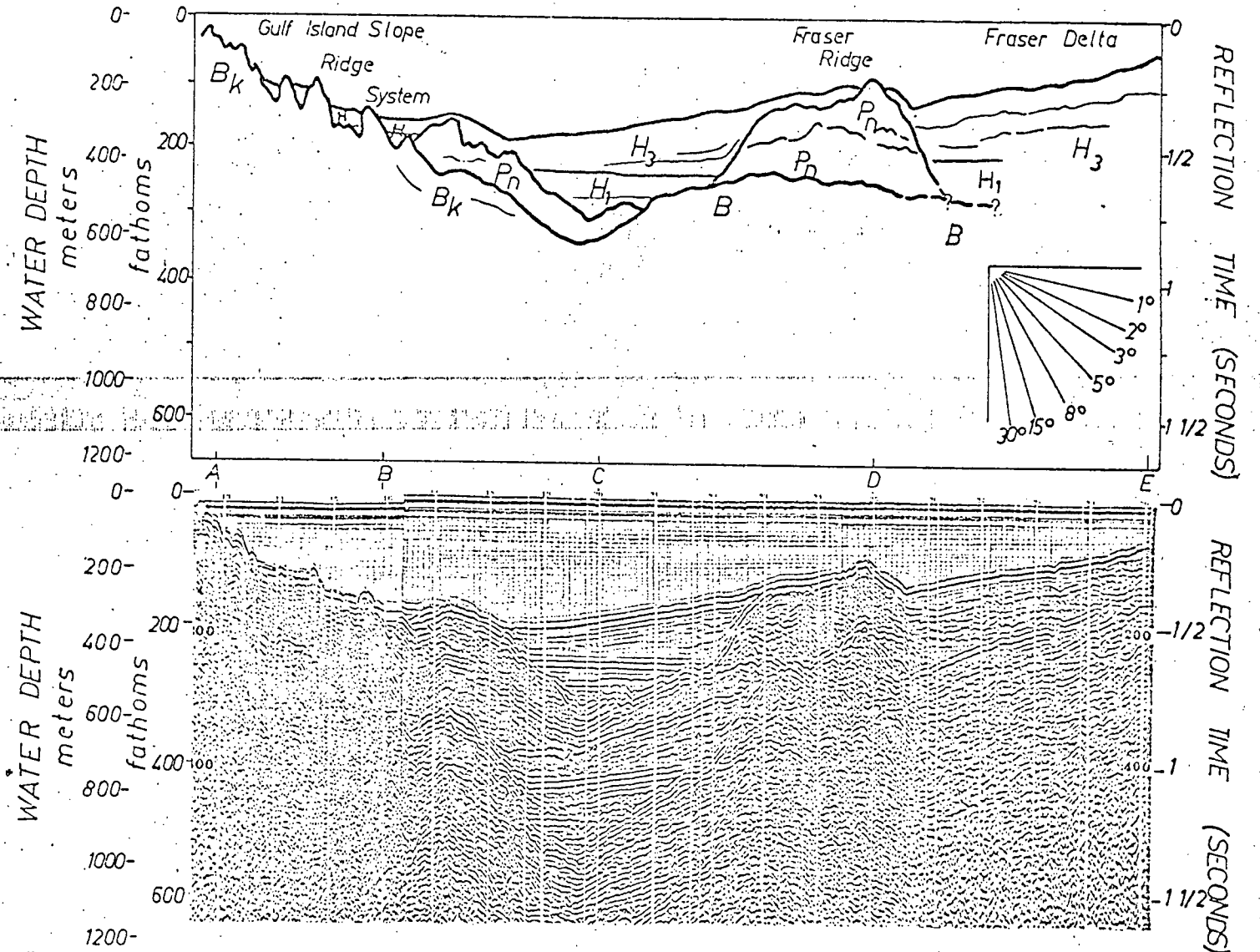
PLATE XVII

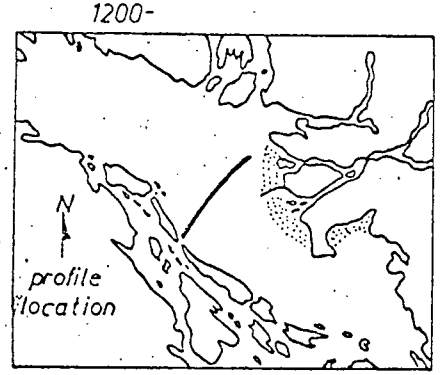




SPARKER PROFILE BY
SHELL CANADA LIMITED
LENGTH 29 KILOMETERS
POSITIONS ON FIGURE 5
LEGEND OPPOSITE PLATE I

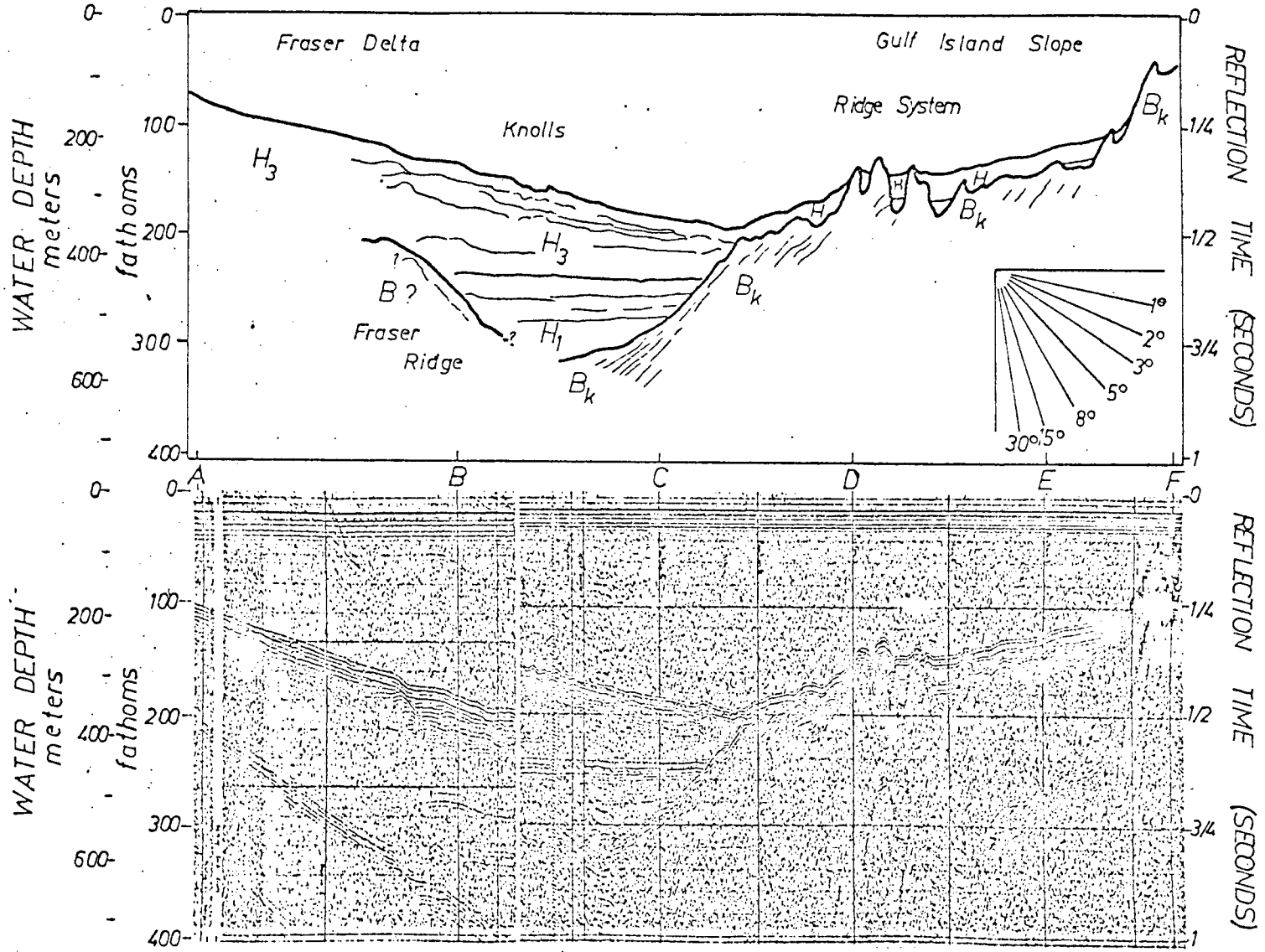
PLATE XVIII

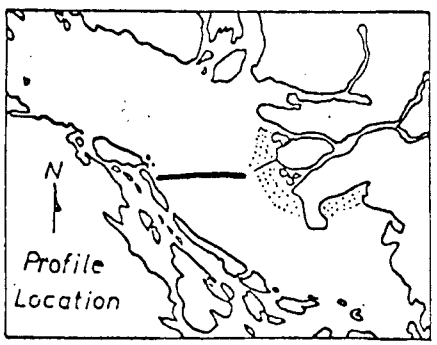




GAS EXPLODER PROFILE BY SHELL CANADA LIMITED. LENGTH 29 KILOMETERS. POSITIONS ON FIGURE . LEGEND OPPOSITE PLATE 1.

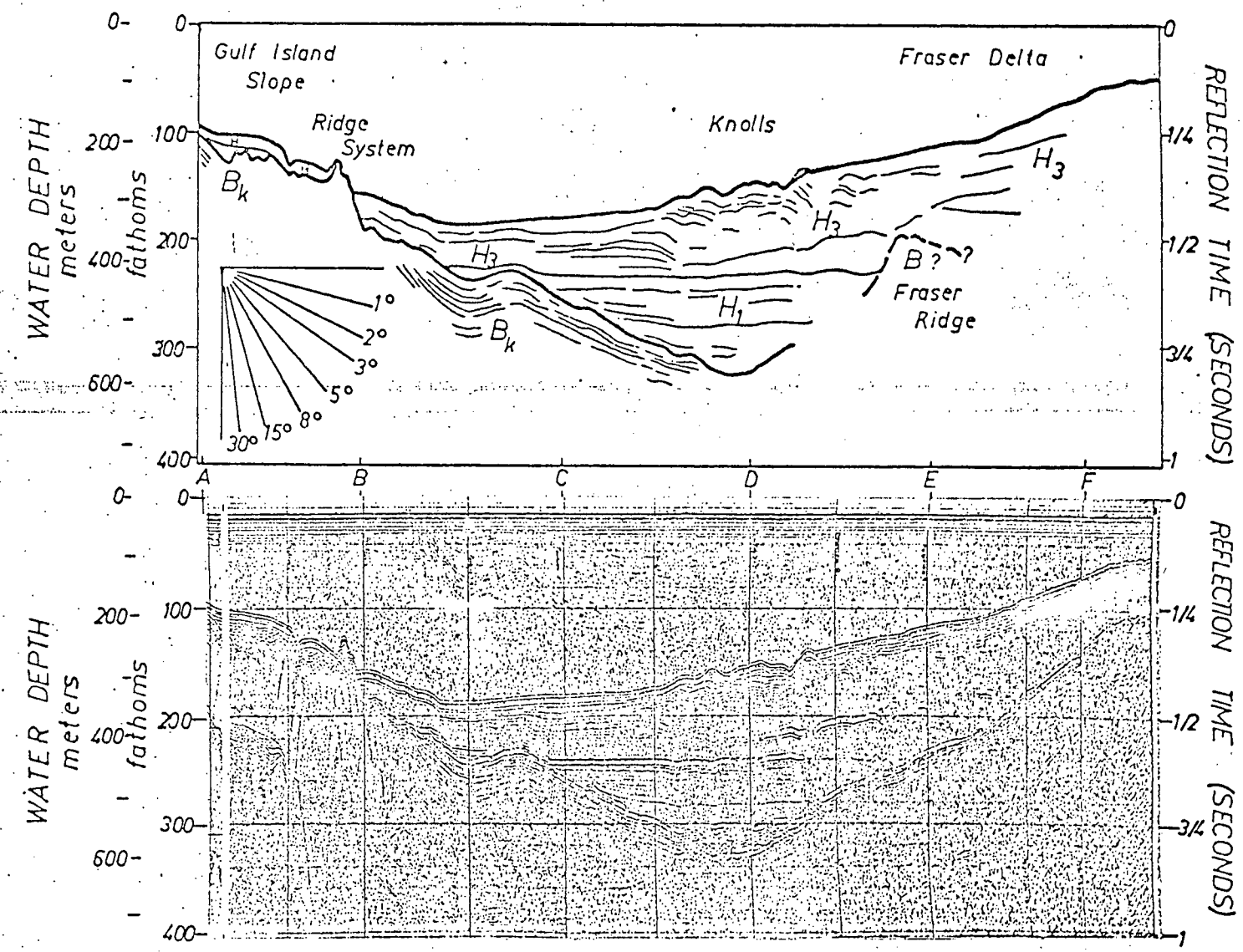
PLATE XIX

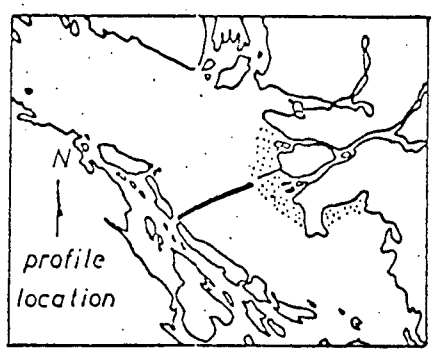




LENGTH 21 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

PLATE XX



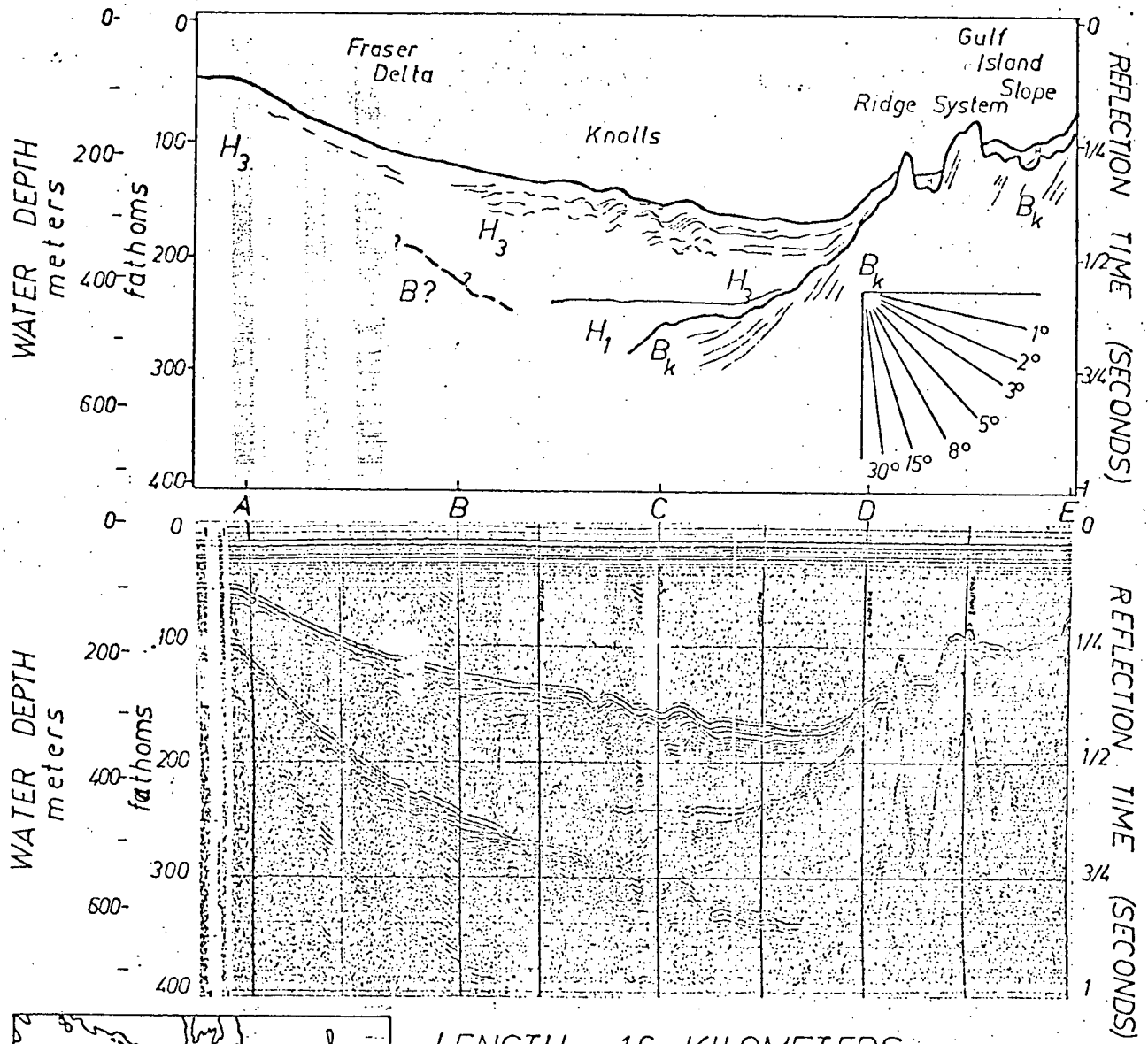


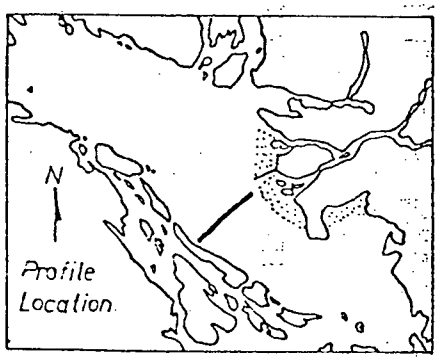
LENGTH 17 KILOMETERS

POSITIONS ON FIGURE 5

LEGEND OPPOSITE PLATE I

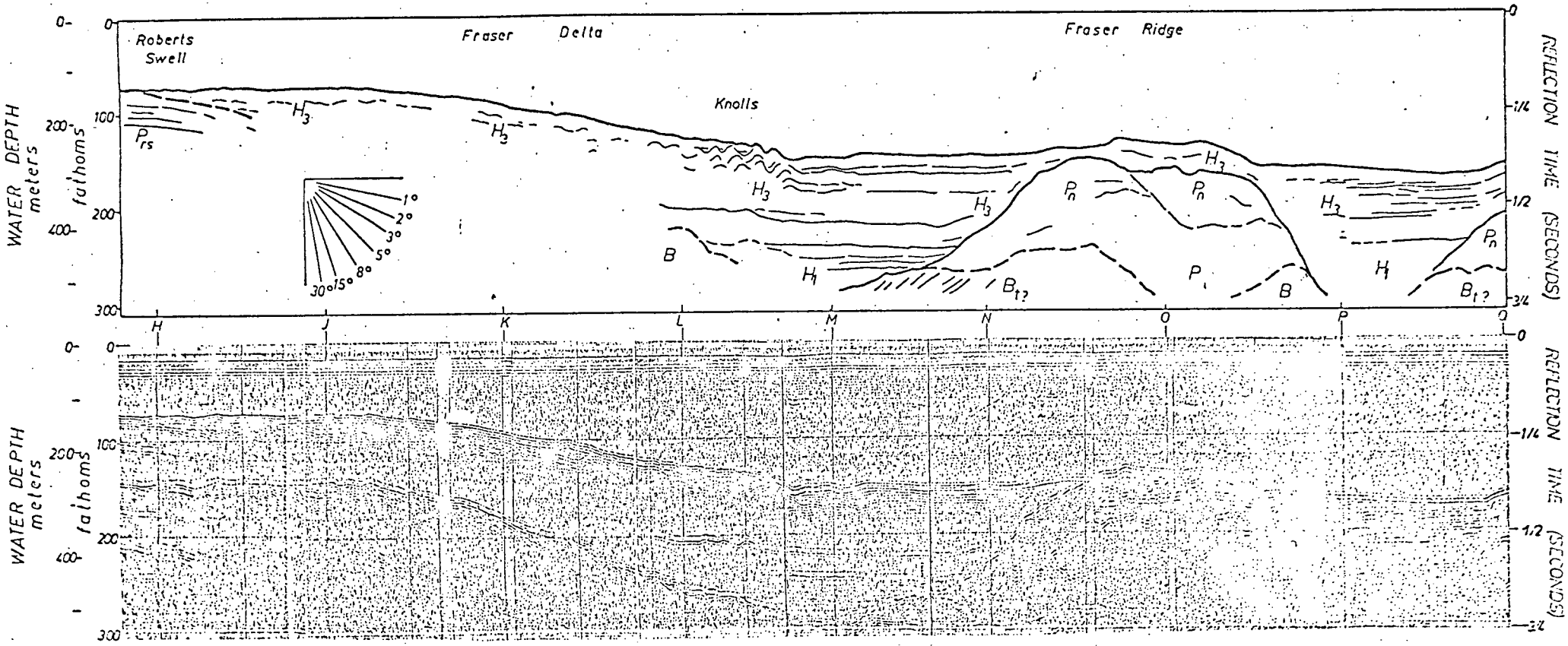
PLATE XXI

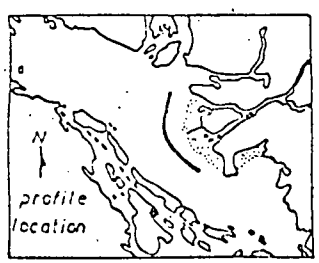




LENGTH 16 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

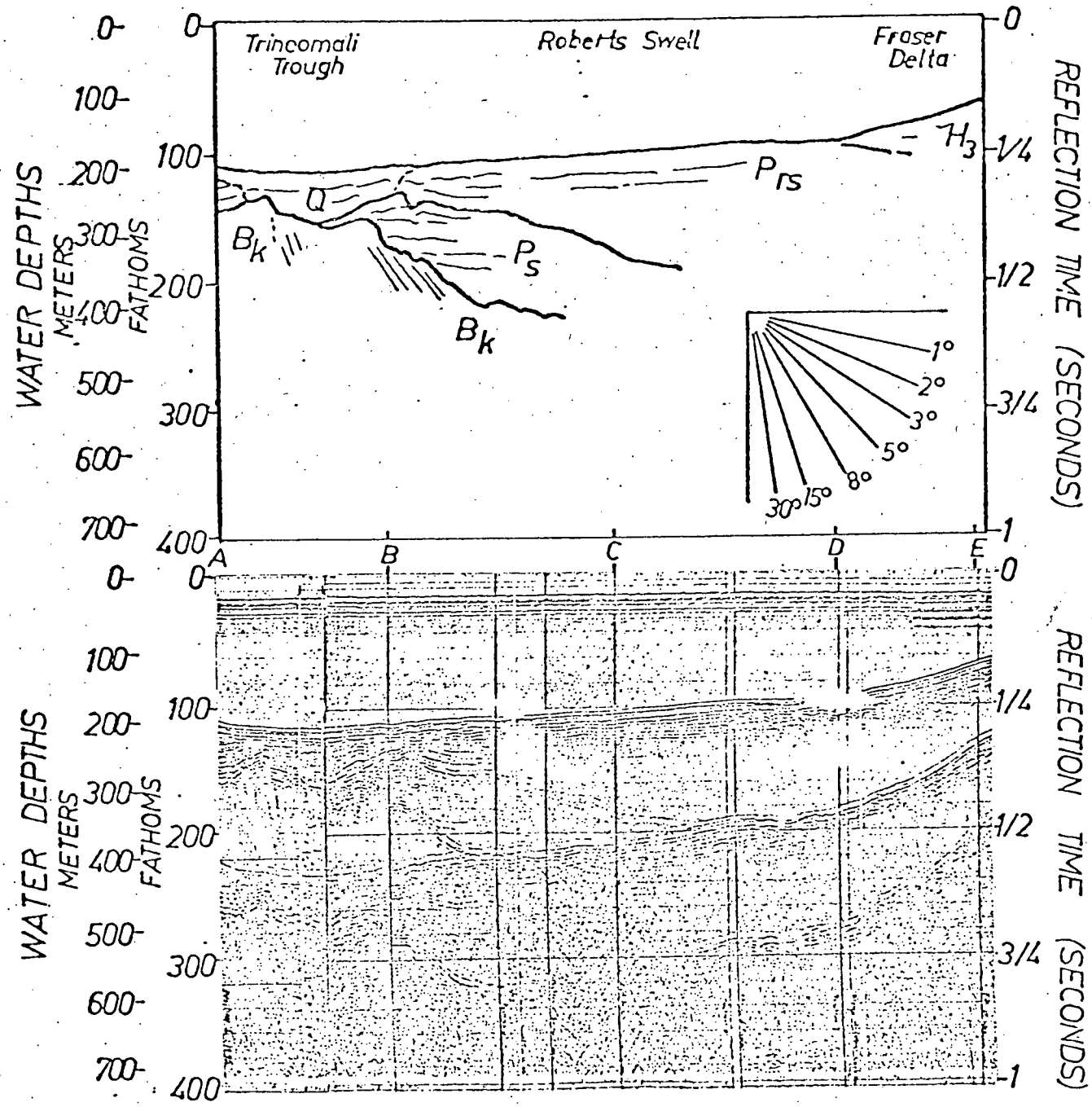
PLATE XXII

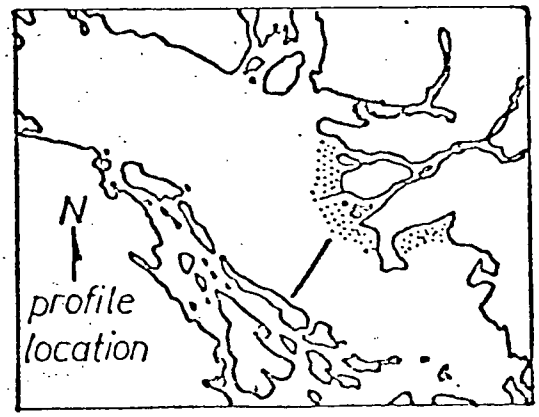




LENGTH 37 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

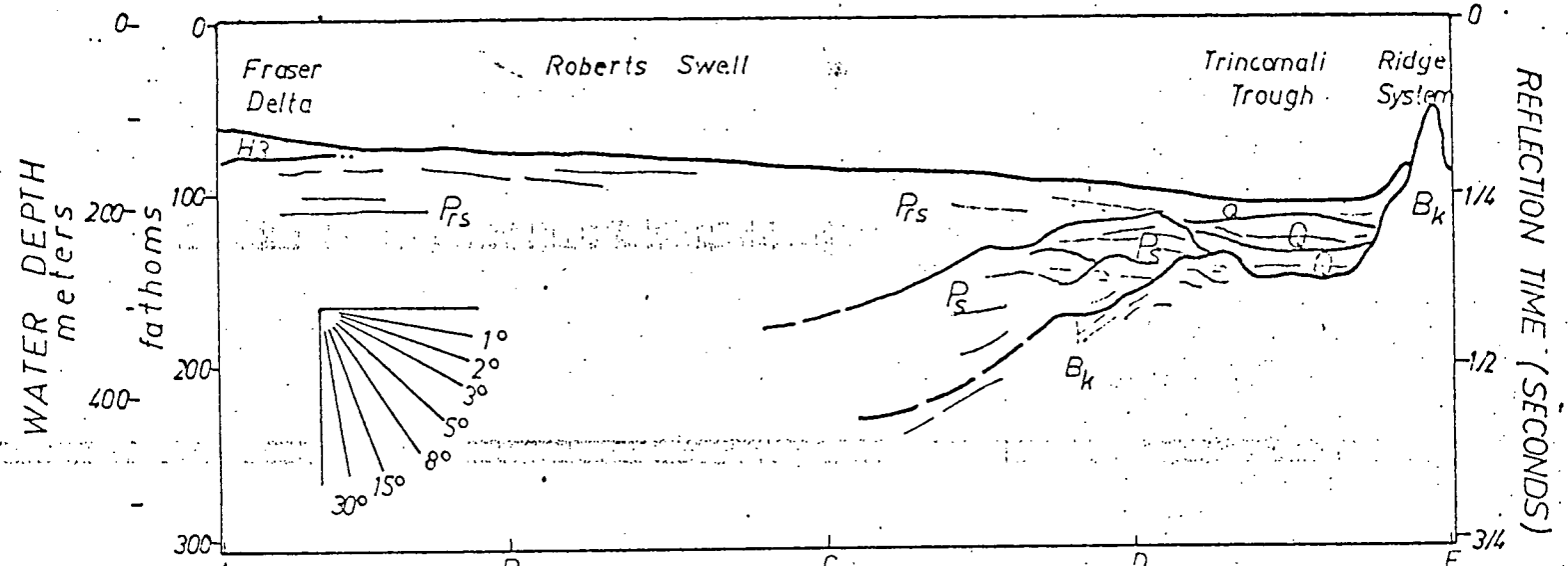
PLATE XXIII

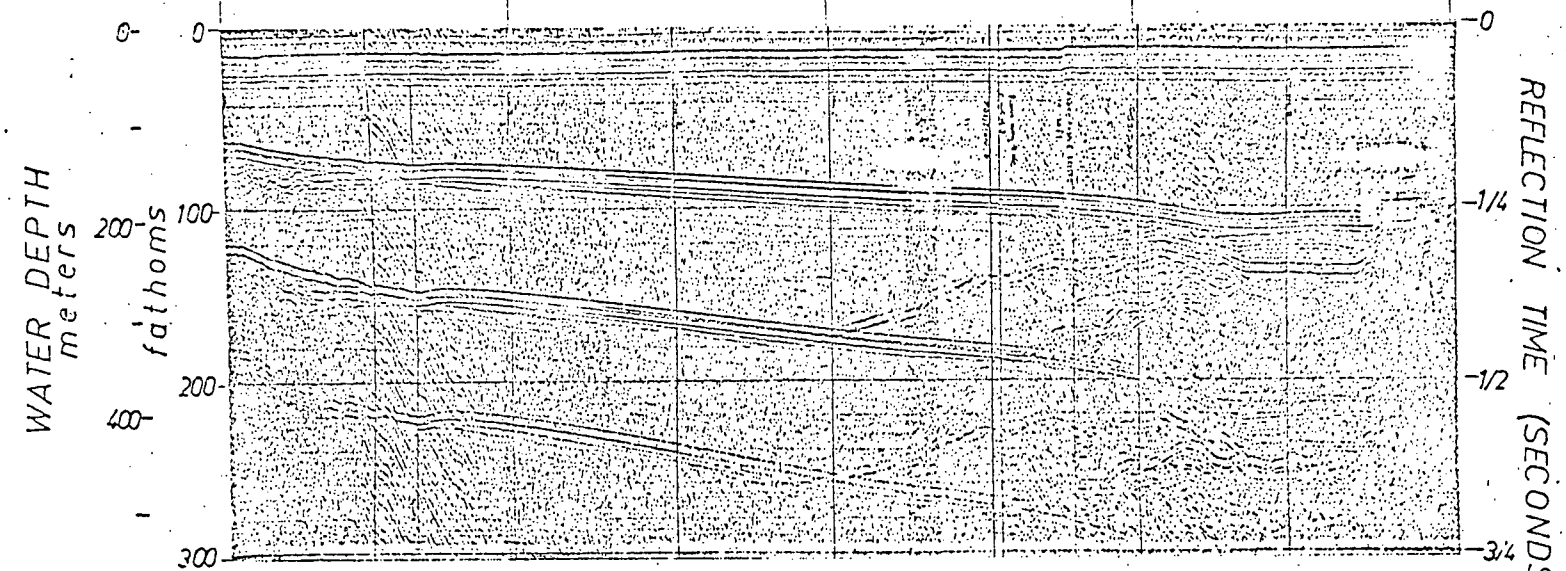


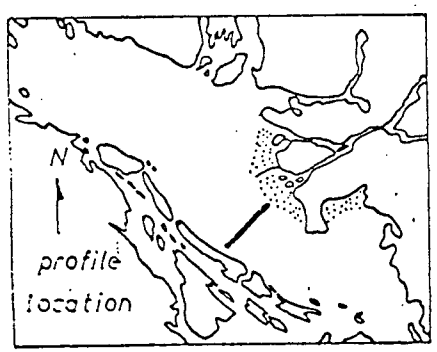


LENGTH 13 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE 1.

PLATE XXIV

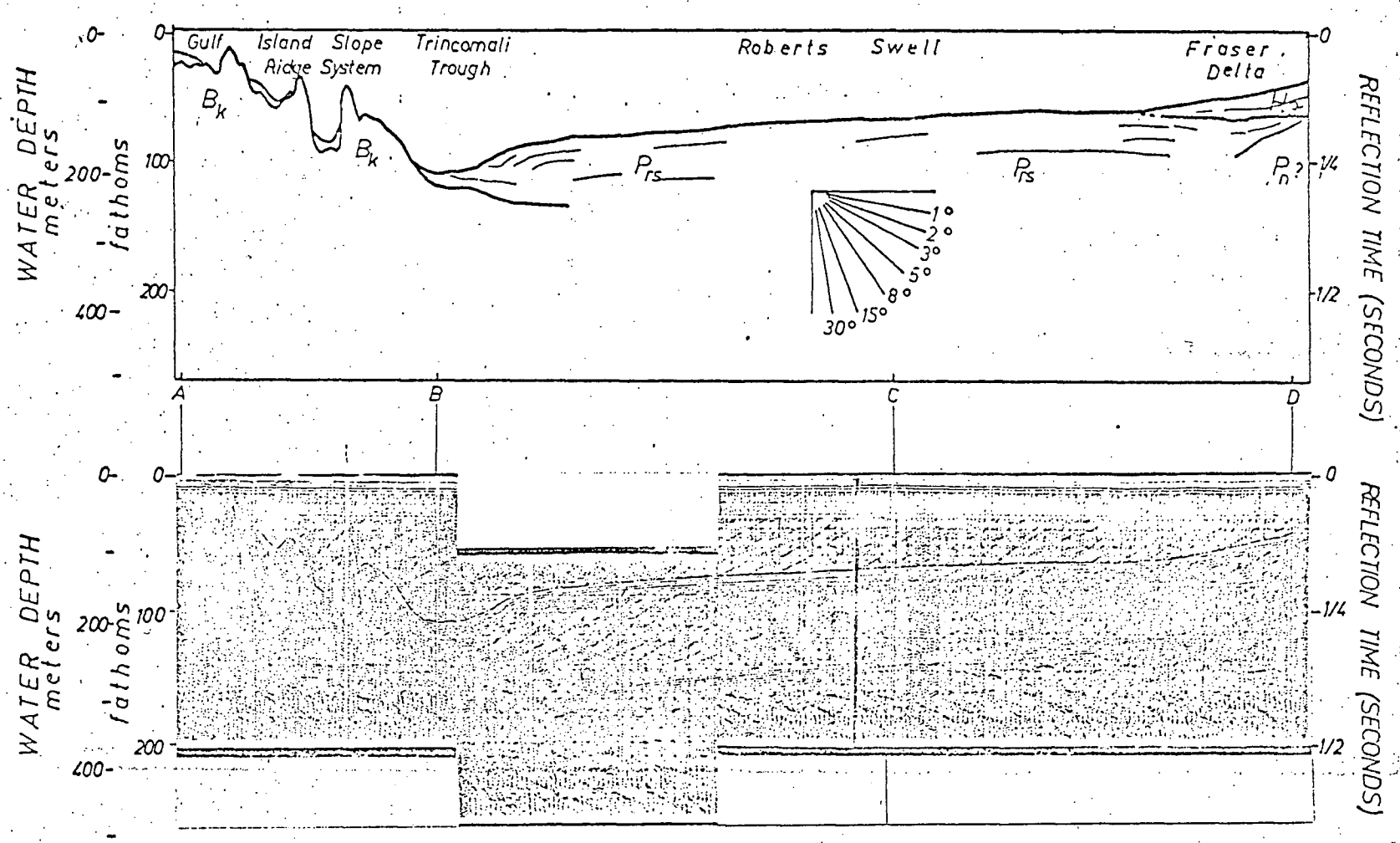


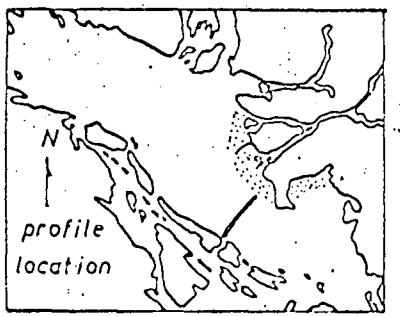




LENGTH 13 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

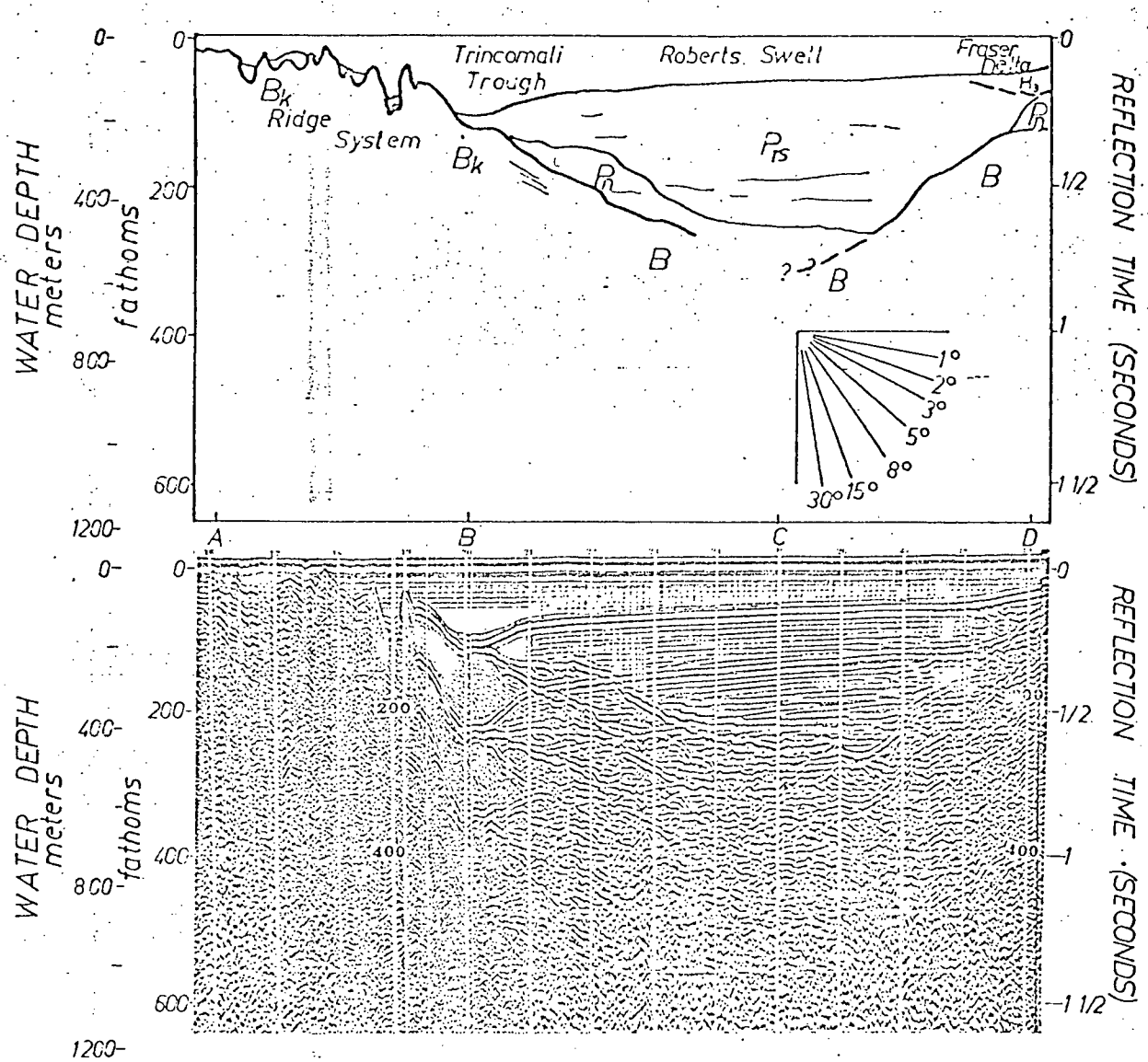
PLATE XXV

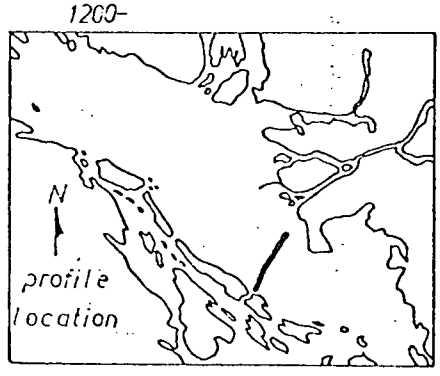




SPARKER PROFILE BY
SHELL CANADA LIMITED
LENGTH 17 KILOMETERS
POSITIONS ON FIGURE
LEGEND. OPPOSITE PLATE 1

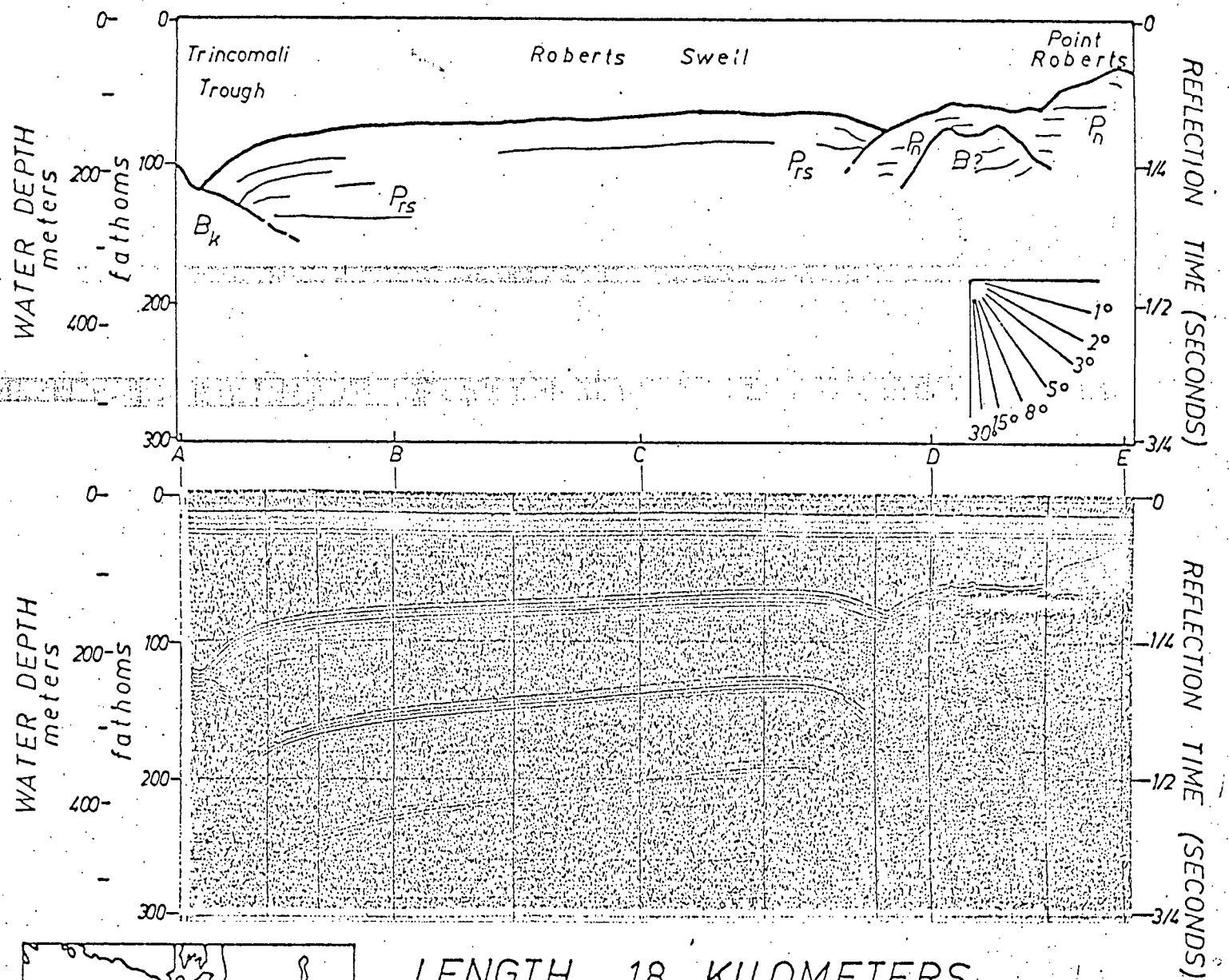
PLATE XXVI

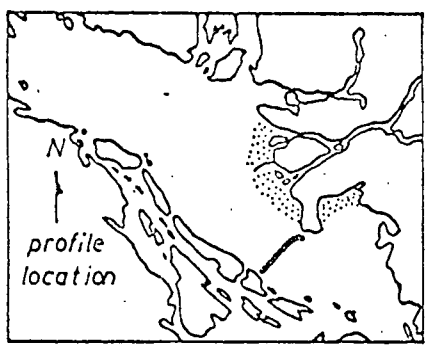




GAS EXPLODER PROFILE BY
SHELL CANADA LIMITED
LENGTH 17 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

PLATE XXVII



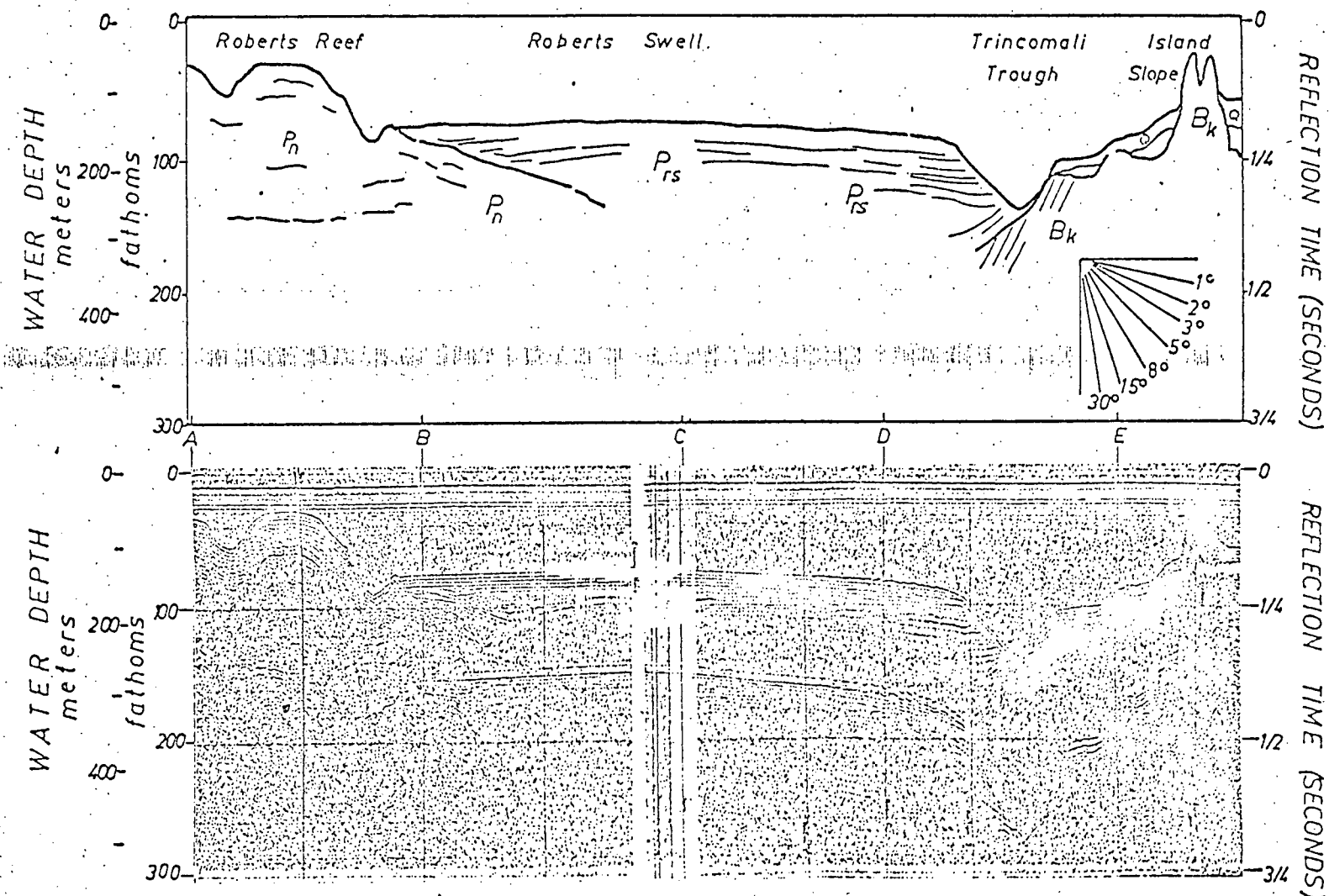


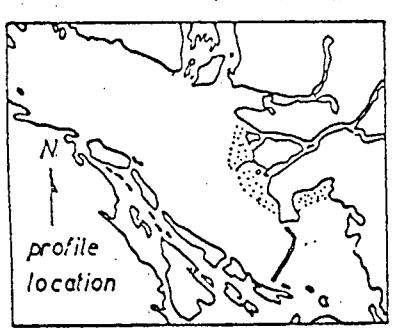
LENGTH 18 KILOMETERS

POSITIONS ON FIGURE 5

LEGEND OPPOSITE PLATE I

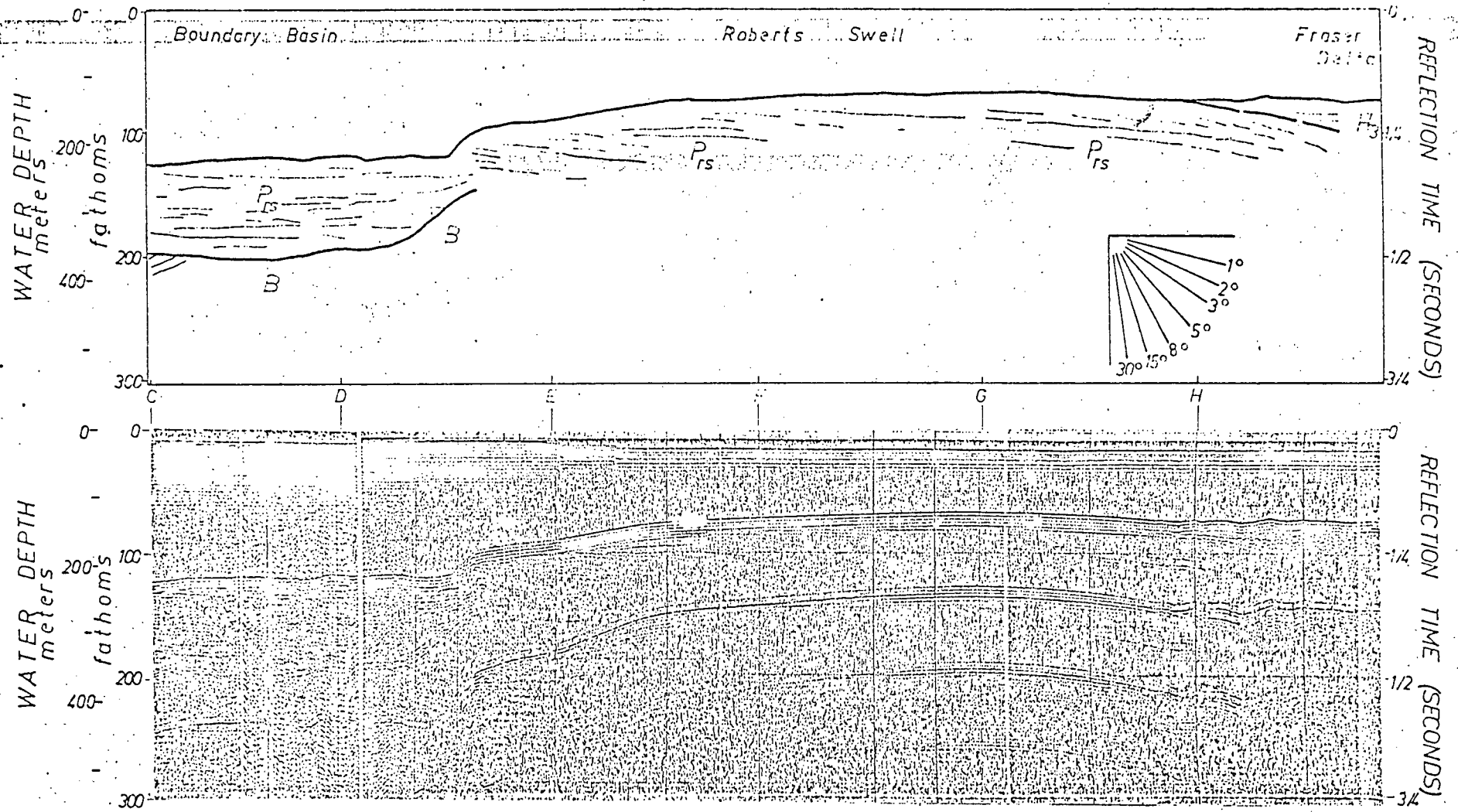
PLATE XXVIII

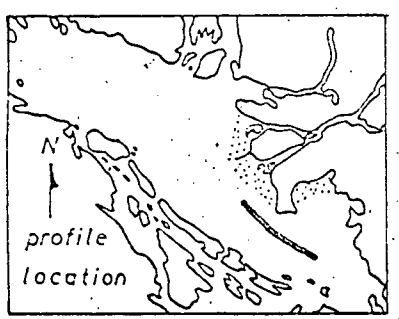




LENGTH 20 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

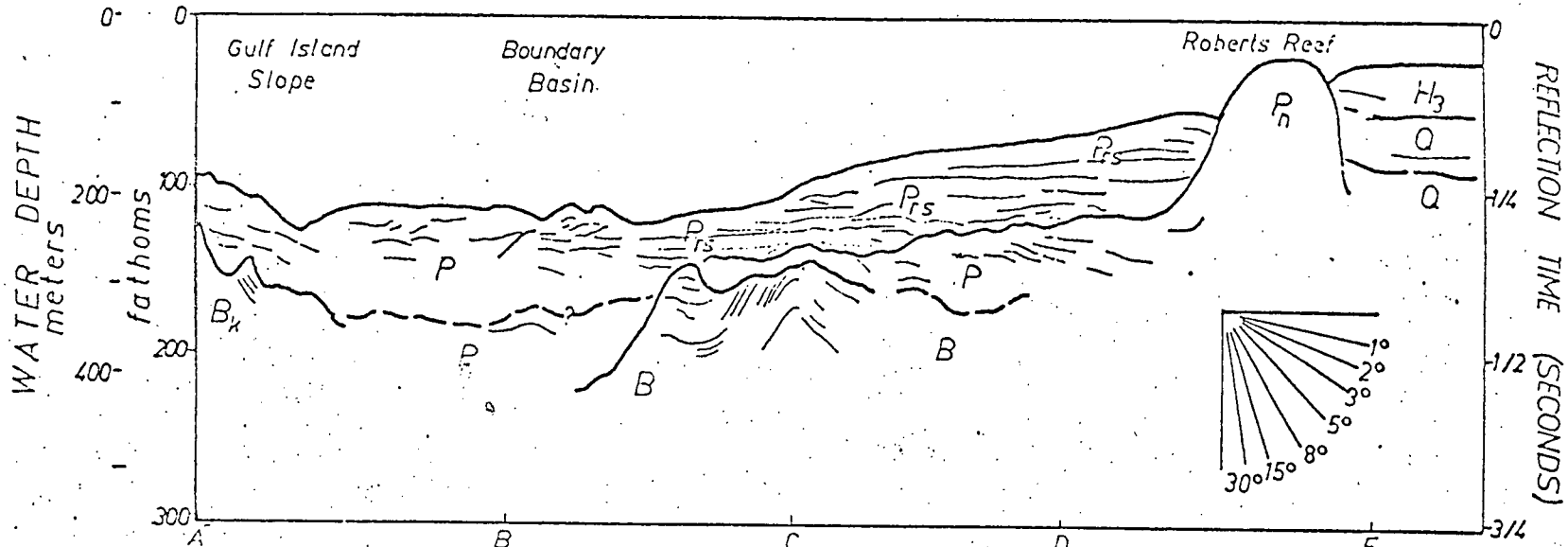
PLATE XXIX

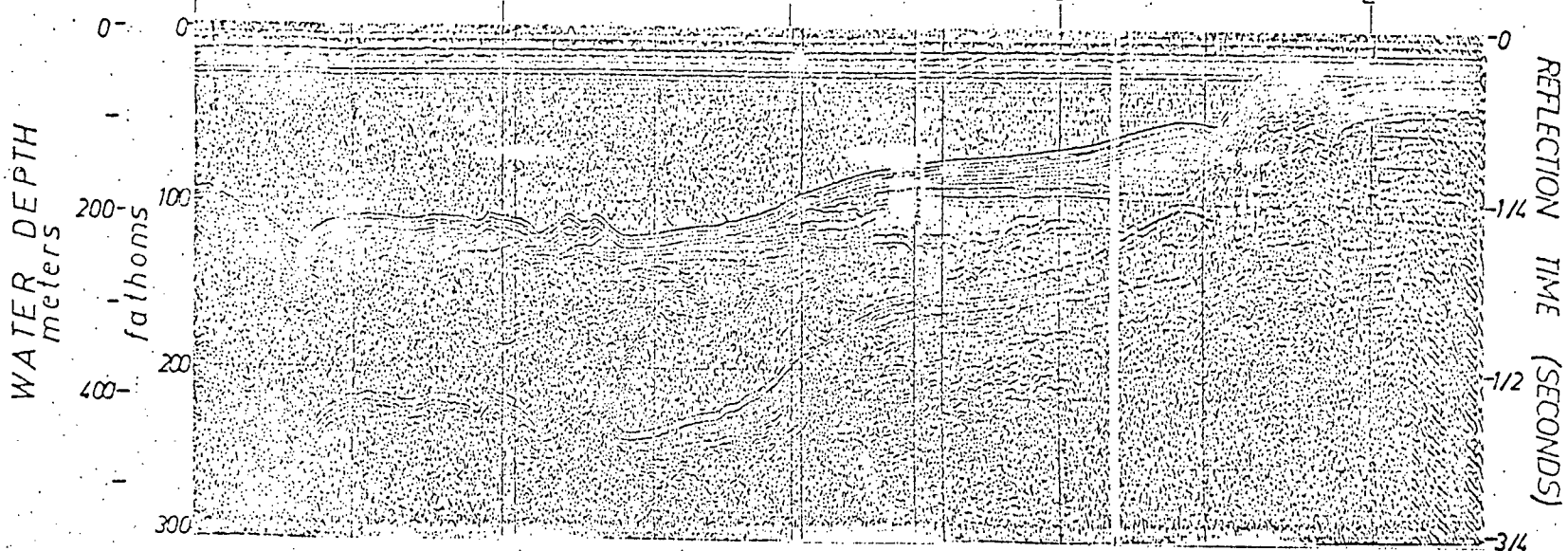


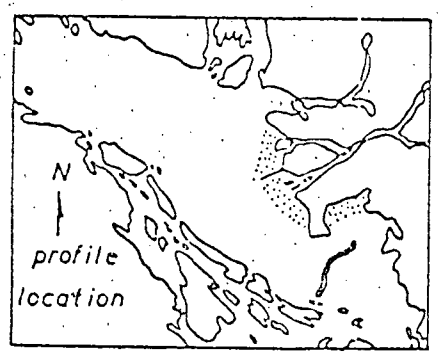


LENGTH 27 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

PLATE XXX





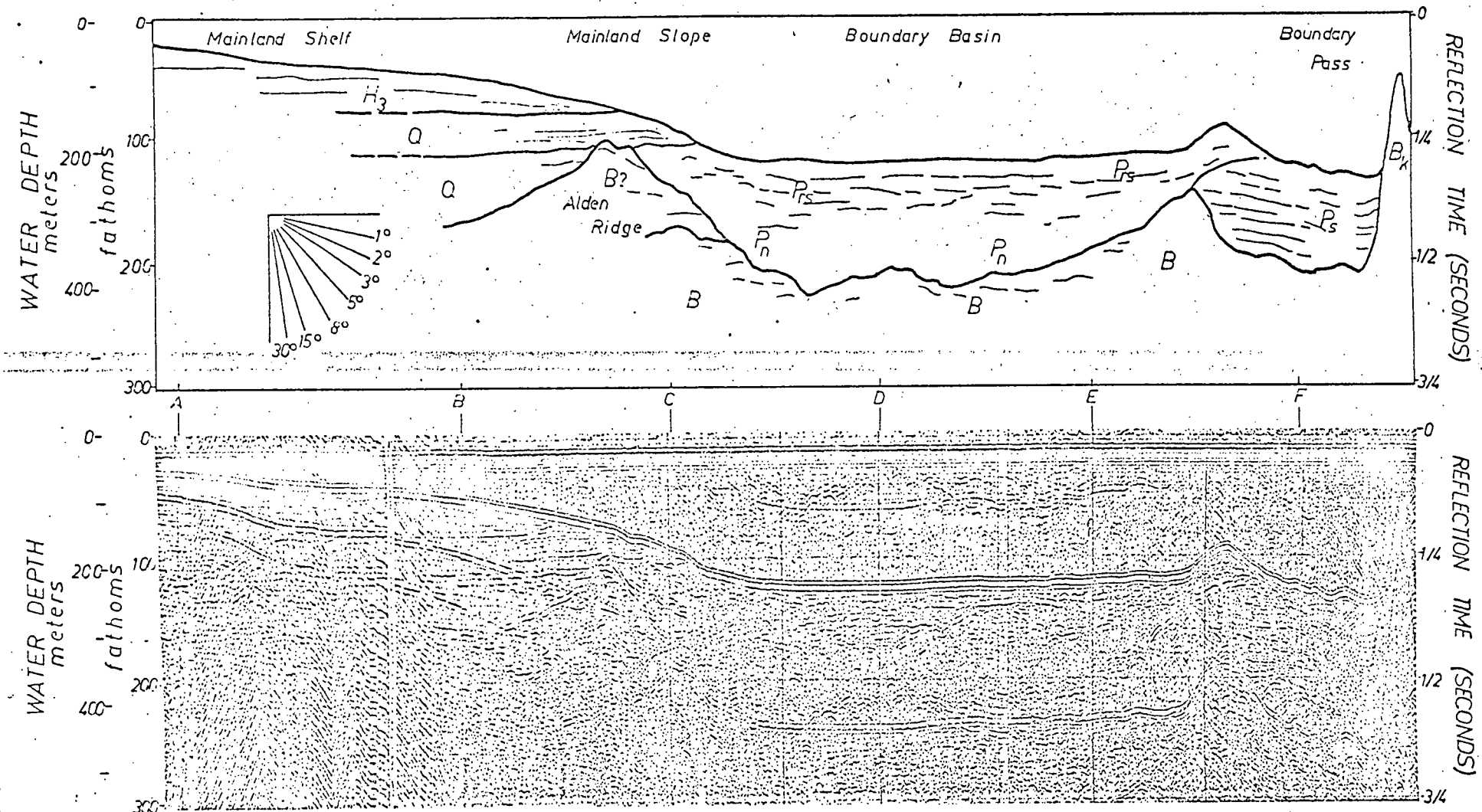


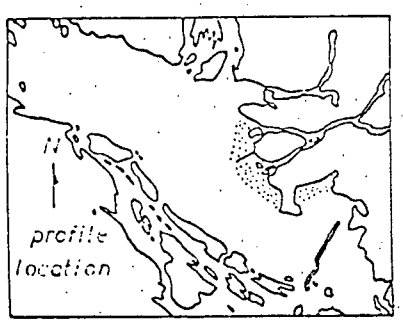
LENGTH 22 KILOMETERS

POSITIONS ON FIGURE

LEGEND OPPOSITE PLATE I

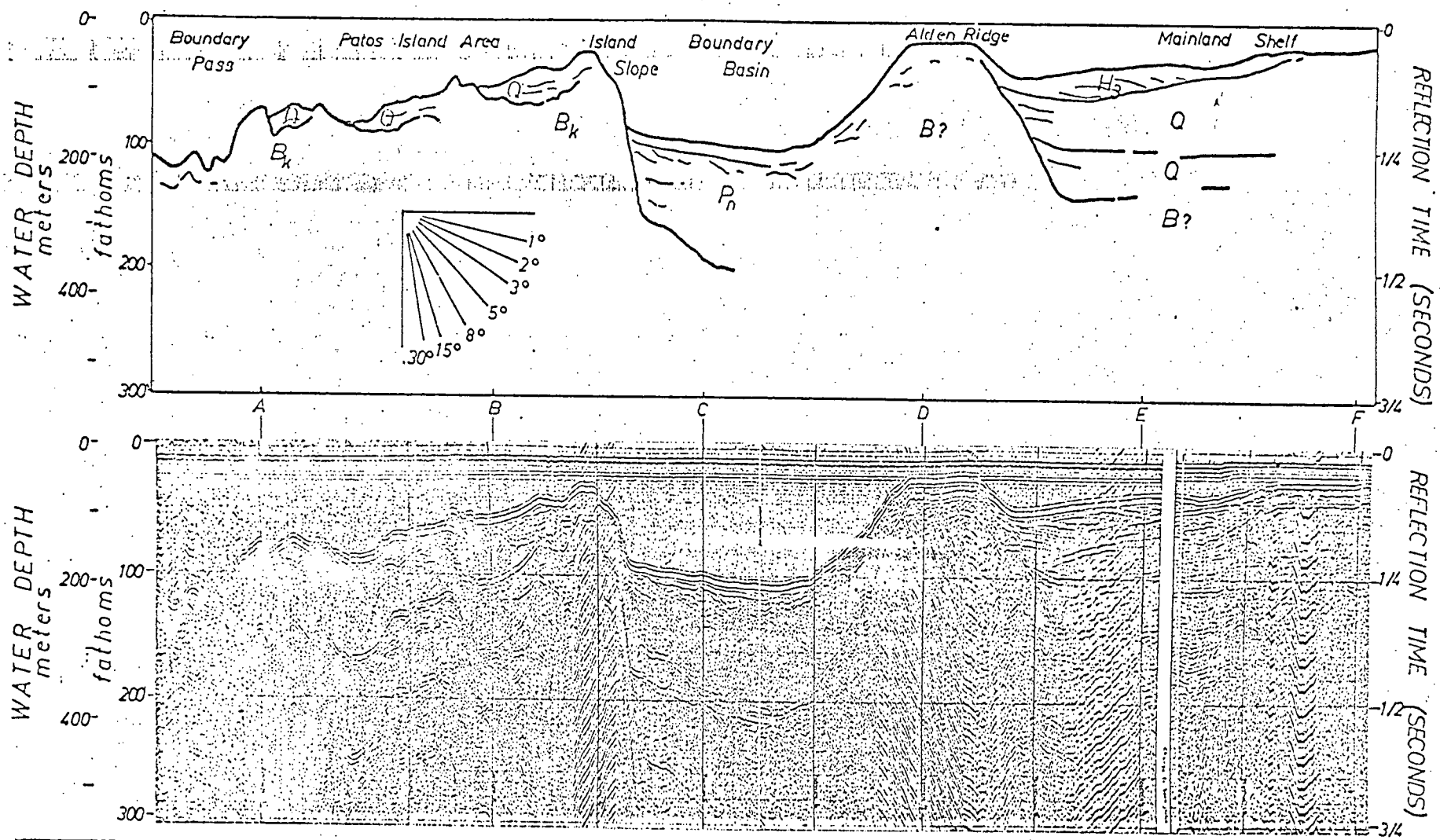
PLATE XXXI





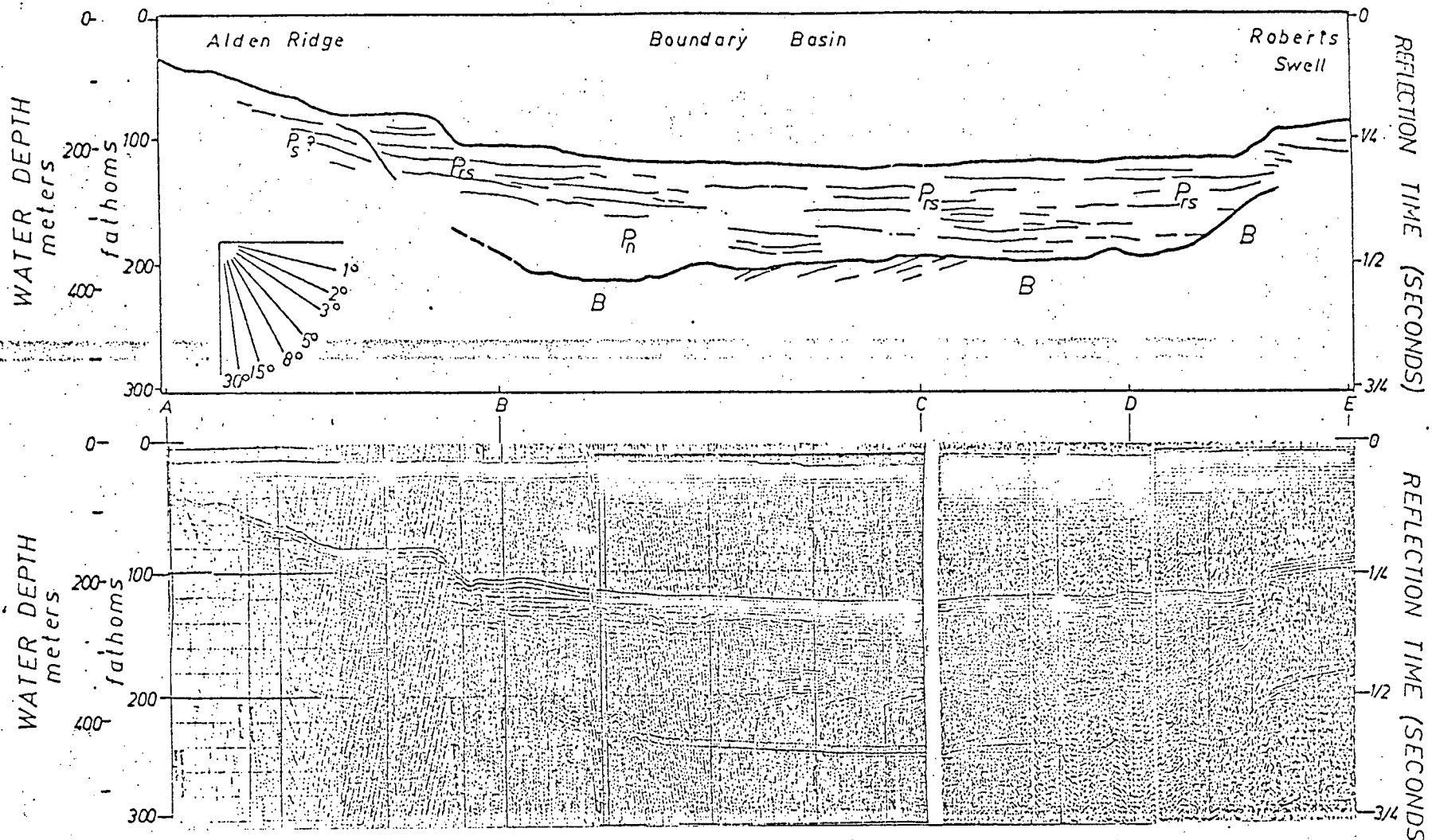
LENGTH 20 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

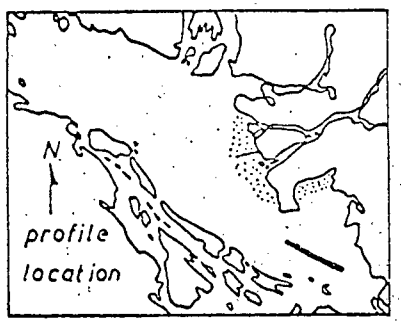
PLATE XXXII



LENGTH 19 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE

PLATE XXXIII





LENGTH 15 KILOMETERS
POSITIONS ON FIGURE
LEGEND OPPOSITE PLATE I

PLATE XXXIV